TREES AND SNOW: THE DEPOSITION OF SNOW ON THE GROUND

A Review And Quantitative Synthesis

INTEGRATED WILDLIFE INTENSIVE FORESTRY RESEARCH

Province of British Columbia

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TREES AND SNOW: THE DEPOSITION OF SNOW ON THE GROUND

A Review and Quantitative Synthesis

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Then the Lord answered Job out of
the whirlwind, and said,
Hath the rain a father? Or who hath
begotten the drops of dew?
Out of whose womb came the ice?
and the hoary frost of heaven, who hath
engendered it?

Job 38, 1, 28-29

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I. THE APPROACH TAKEN

1. INTRODUCTION

Miller (1966: 1) observed "Forest covers much of that portion of the earth that receives some of its precipitation in the form of snowfall. Yet many questions about the amount of snowfall that penetrates forests to reach the soil surface remain unanswered." Nearly two decades later Miller's statement remains true. The effects of forest on snow cover are variable. Seppanen (1961) reported that in some places in Finland, more snow accumulated in forest than in open locations, and in other places less; the relationship differed between years. Harestad (1979) documented the same phenomenon on Vancouver Island.

In many areas characteristics of the snowpack influence mortality rates and growth rates within wildlife populations (e.g., Formozov 1946, Severinghaus 1947, Nasimovich 1955, Jones and Bunnell 1985). Any environmental feature that can modify snow characteristics can modify the influence of these characteristics upon wildlife. The environmental feature over which man exerts greatest control is vegetative cover; specifically, forest cover. Furthermore, there is abundant evidence that forest cover does modify the influence of snow upon wildlife (Shank and Bunnell 1982a and b). Given these broad relationships between trees, snow, and wildlife, it is apparent that man can influence wildlife populations by the way in which he modifies forest cover. One important set of influences involves the manner in which forest cover influences snow deposition on the ground. That broad area is the subject of this review.

The potential for snow characteristics to influence wildlife populations exists over much of British Columbia. The potential for man to modify snow characteristics by manipulating forest cover is also widely spread. This review has a more specific geographic focus; namely, southwestern British Columbia. Four compelling reasons encourage a specific geographic focus:

1) There are ample data documenting that during winters with prolonged, deep snowpacks, coastal black-tailed deer do best in old-growth forest cover (e.g., Bunnell 1979, Hebert 1979, Harestad et al. 1982, Jones and Bunnell 1985, Bunnell and Jones 1985). However, reservation of old-growth forests to provide winter range is in direct conflict with maximum wood-fibre production.

2) Southwestern British Columbia contains the most productive areas for both deer and commercial forests in the province, often on the same land (e.g., Hebert 1979, Bunnell 1983, B.C. Ministry of Environment and B.C. Ministry of Forests 1983). That phenomenon intensifies potential conflict between resource uses.

3) The presence of the Integrated Wildlife and Intensive Forestry Research Program (IWIFR) in southwestern B.C., provides a useful vehicle for examining forestry-wildlife relationships and for incorporating these findings into management there.

4) Generally, less is known about coastal snow characteristics than is known about more continental regions. The information that is relevant to southwestern B.C. is widely scattered and requires collation.

In summary, although many of the relationships addressed in this report occur throughout north temperate regions the review has concentrated on those most applicable to southwestern B.C. because it is there that conflicts involving relationships between trees, snow, and wildlife are most intense. The broadest objective of the review is:

> to summarize and quantitatively analyze those data relating to delivery of snow and the influence of forest cover on snow deposition on the ground.

The summary is meant to facilitate more efficient research and management activities directed to:

1) understanding how old-growth forests produce snow characteristics more favourable to deer than do younger

stands, and

2) creating effective winter ranges in managed stands.

1.1 Scope

The geographic focus has been noted. Because concerns for black-tailed deer populations were a major impetus for this review, snow characteristics initially considered were those believed important to deer. They include: i) depth; ii) density; iii) hardness; iv) spatial distribution; and v) temporal duration.

These characteristics, almost certainly, will be important to other ungulates such as caribou, moose, or elk (Bunnell 1978, Harestad and Bunnell 1979). Likewise they will influence smaller wildlife species utilizing the nivean or sub-nivean environment during winter (Formozov 1946). The characteristics are also relevant to hydrological studies. The manner in which all these characteristics influence wildlife is not addressed in this review. The review is limited to relationships between forest cover and the manner in which snow is deposited on the ground. It treats only snow depth and spatial distribution in detail.

Given one of the intended purposes of the review (provision of information relevant to the production of winter ranges in managed stands), the review is process-oriented. There are two broad reasons for the process orientation.

First, Miller's (1966) statement cited at the beginning remains true largely because individual processes have been examined only infrequently. The diversity of forms assumed by trees and stands, the diversity of forms of snowfall and conditions under which it is delivered to forests, and the inadequacy of sampling and measuring techniques make it impracticable to settle questions by direct measurement. Furthermore, simple but universally applicable relationships between snowfall in forested and open areas that might serve purposes of prediction have proven elusive. Both the complexity and the failure to extract simple relationships lead us to conclude (as did Miller 1966) that the manner by which forests receive, intercept, hold, and dispose of snowfall is not a simple unitary phenomenon but a series of events or processes which require independent study. Second, we believe that the kinds of stand structure or silvicultural treatments most likely to simulate ideal winter range conditions may not have been attempted yet. Simple documentation of empirical findings would prove insufficient to predict results of untried treatments. If prediction of the effects of untried treatments is important, it is necessary to attempt to extricate and document individual processes relating forest characteristics to snow characteristics.

Similarly, it is important to document boundary conditions where one process becomes ineffective or is dominated by another. If the approach were completely successful, it would allow an understanding of the processes that would permit the manager to match a specific forest structure with target snow conditions. That approach necessitates the erection of a complete and internally consistent, conceptual model of individual processes. The notions of individual processes and boundary conditions necessarily expand the scope. For example, to document the nature of a process one must examine it at a finer level of detail than merely describing the effects it has, especially when that process and several other processes are acting simultaneously. The specific nature of the approach is treated in greater detail later (Ch. 1.3). Here we note two broad points relevant to scope of the review:

1) The conceptual model with its orientation to specific processes and boundary conditions has determined the scope. All processes or fixed environmental features believed important in determining the pattern of snow deposition on the ground are addressed.

2) The specific literature synthesized is generally a function of the detail with which any particular process is treated and the anticipated geographic variability of that process. For example, treatment of air masses and storm tracks is largely restricted to literature treating southwestern British Columbia. These two phenomena are treated at a coarse level of resolution and exhibit great geographic variability. Conversely, treatment of

processes at a fine level of resolution (e.g., growth rates of ice crystals or effects of angle of the intercepting surface on snow accumulation) utilize much of the available data, regardless of the study location. It is assumed that at a fine level of resolution processes differ little geographically, providing the important environmental conditions are acknowledged. The latter proviso emphasizes the importance of an appropriate conceptual framework.

1.2 Objectives

The broadest objective is the provision of organized information in a fashion that will effectively focus research on tree:snow relationships and will assist attempts at production of winter range in managed stands. From this overriding objective several specific objectives follow. These include:

1) To provide a conceptual framework or word model that describes the manner in which forests modify the deposition of snow.

2) To extract and describe the separate processes underlying patterns of snow deposition.

3) To describe boundary conditions that determine the

effective influence of individual processes affecting snow deposition.

4) To quantify individual processes and relationships with the goal of providing useful predictive relationships.

5) To document the particular processes or kinds of processes most likely to be important in southwestern B.C.

6) To evaluate whether relationships observed elsewhere are applicable to southwestern B.C.

1.3 The Approach

The broad approach taken was introduced earlier to indicate how the approach has influenced the scope (Section 1.1). Briefly, to anticipate effects of stand treatments not yet applied one must extrapolate current knowledge. Effective extrapolation of current, empirical, and integrative relationships requires understanding of the underlying processes. Ideally one must separate many intermingled phenomena to arrive at an understanding of any one. Unfortunately, the effect of forests on snow is not a single problem, but many. The forest acts on aerodynamic processes that govern the deposition of snow. It also intercepts some fraction of the snow. It acts, in a different manner, on transfer of heat that melts the snow. Furthermore, it acts differently on every avenue of heat exchange - short-wave radiation, long-wave radiation, convection, conduction, evaporation, and condensation (Fig. 1.1). Not all characteristics of snow considered important are illustrated in Figure 1.1, neither are those relationships which are illustrated presented completely. The figure serves primarily to reveal a portion of the complexity.

To fulfill the objective of describing and quantifying individual processes requires a conceptual model, of which some portions can not be addressed adequately. There are three main obstacles to the complete model. One is that most data available are inappropriate to the scale of interest - that of a manager manipulating individual stands. Meteorologists and hydrologists most often work at a much larger scale; the manner in which they define, measure, and describe processes reflects that larger scale. On the other hand, the micrometeorologists and cloud physicists work at a much finer scale, defining and describing processes at that level. There is seldom a convenient and effective way to join these disparate levels of resolution. A second, more intractable problem, is that many interactions occur simultaneously. All four major modes of heat exchange (processes) may be occurring at one time, each influenced differently by forest cover. Most literature has evaded the problem by not attempting separation. One result is the inability to define what has been measured. That inability impedes synthesis because dimensions, along which individual

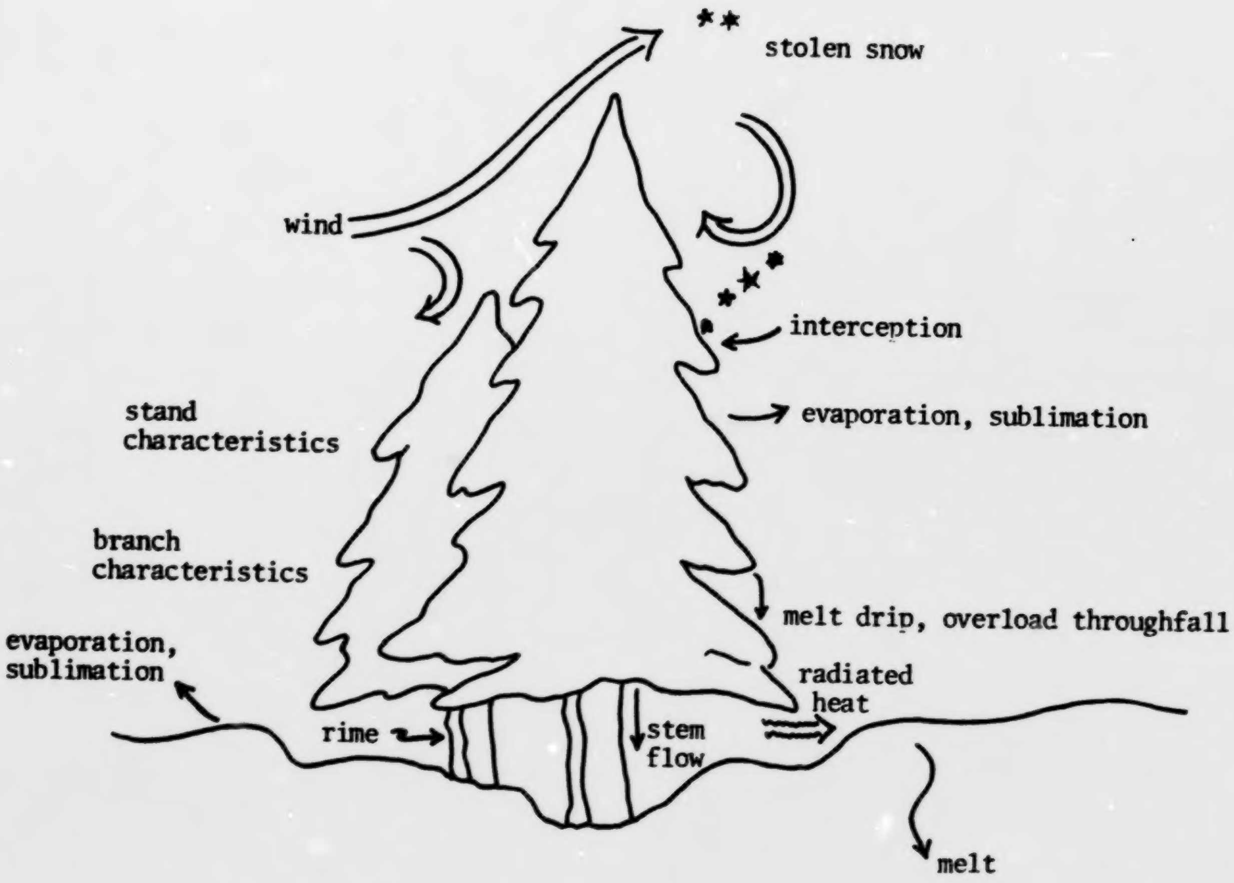
stolen snow
wind
interception
evaporation, sublimation
stand characteristics
branch characteristics
melt drip, overload throughfall
evaporation, sublimation
radiated heat
rime
stem flow
melt

Figure 1.1 Incomplete schematic representation of relationships governing observed interception of snow by forests.

studies might be arrayed to reveal general patterns, are ill-defined. The third difficulty is in part a product of the first two and in part a common phenomenon of large systems: the whole is qualitatively more than the sum of the parts. The something more is conventionally termed an emergent quality (see Churchman 1968). One example is that one cannot simply sum the processes documented for individual trees to predict interception by a stand. Qualitatively different patterns occur (compare Chapters 7 and 8).

Despite these obstacles we have attempted to pursue individual processes wherever possible. The conceptual model represents our attempts to bridge disparate levels of resolution and to indicate how individual processes may interact. In several instances we have had to reanalyze published data to extract particular relationships that were important within the conceptual model (e.g., Table 3.2, Figs. 3.22, 7.36-7.44). Sometimes we have had to develop new definitions and approaches to describe and predict functional relationships (e.g., Chs. 6 and 7, Harestad and Bunnell 1981). Our intent has been to extract predictive relationships such as regression models wherever possible. Statistical analyses have thus attempted to document which environmental variable was the most efficient predictor of a given snow characteristic (lowest standard error of regression). About 25 years ago, Goodell (1959) discussing the influences of forests on snowpacks, noted that there was a deficiency in "basic data to test hypotheses" and of suitable instruments.

Instrumentation has improved, but it is still not always possible to provide a sound statistical relationship; in such instances, a verbal description of the process is presented. The entire report is best described as a quantitative 'word model' discussing relationships between trees and snow.

Before beginning the word model itself three points about the approach should be noted explicitly.

1) Processes and observations. This point has been addressed. Briefly, we are attempting to document underlying processes from available observations. Many of the data used were not collected to document a process, but to describe a broader empirical relationship. Some analyses thus fall short of what should be possible with appropriate data.

2) Means and variability. Current models relating deer use or energetics to snow depth or density implicitly assume that only mean snow depths or densities are important (e.g., Harestad et al. 1982, Bunnell and Jones 1985). No consideration is given to spatial variability. We cannot ignore the fact that there is an important trade-off between mean depth and patchiness of distribution. For example, we may expect more deer in an area of which one half is bare and one half covered with 80 cm of snow, than in an area the same size but uniformly covered by 40 cm of snow.

One apparent solution would be to replace the mean with the variance or the variance to mean ratio. However, because variance is a function of the mean and because observed snow depths cannot be < 0.0, the variance has a maximal value at some moderate snow depth. Continuously increasing the patchiness of snow distribution, at any given snow depth, does not result in a continuous increase in variance (Fig. 1.2).

If the relationship is to be vested with biological meaning to the deer it appears that collapsing the spatial variation and mean depth into a simple, synthetic dimension will be inadequate. Field work evaluating which distributions of snow depths provide favourable winter range will require a multi-dimensional approach. Grids will likely prove more revealing than transects; measures of kurtosis and skewness, more revealing than variances and means. Most published data do not provide variances; none provide measures of kurtosis or skewness.

The word model is a compromise. It treats many processes as occurring at a dimensionless point. That approach is probably justified for processes at a detailed level when primarily physical factors are involved. The approach is inadequate for processes or relationships which have an inherent spatial component. The compromise of the word model is to include measurements of variability where possible.

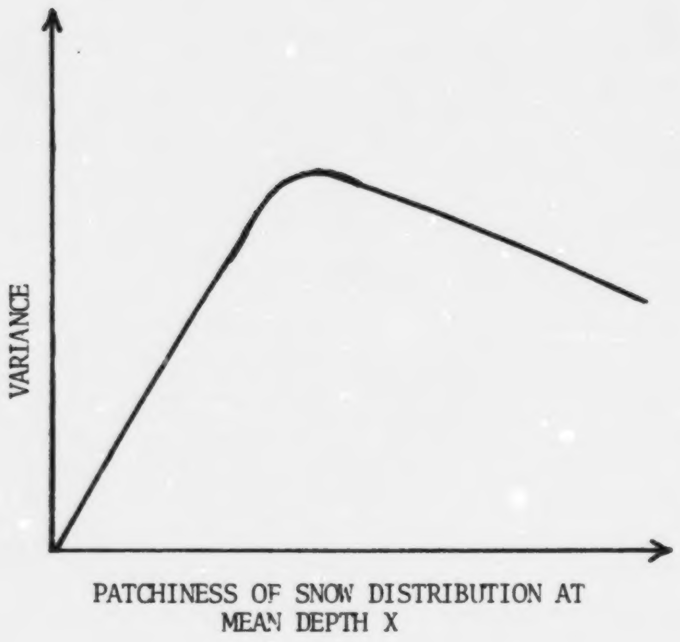

Figure 1.2 Schematic relationship of the change in measured variance as a function of increasing patchiness of snow distribution.

3) Snow events and snowpacks. Snowpacks have a history which is reflected in such hydrological terms as the 'ripeness' or 'maturity' of the snowpack. The state of the snowpack at any time is some integration of past snow events and intervening conditions when snow did not fall (e.g., rain on snow events). Very different histories may produce similar snowpacks. Because the word model attempts to extricate processes, it emphasizes snow events. Data on snowpacks are treated only as an evaluation of the predictive ability of the processes describing and relating snow events. Independent evaluations, not presented in the word model, are being pursued using simulation models (Shank and Bunnell 1982c).

1.4 Organization

The word model treats two broad groups of processes: those affecting delivery of snow to a surface, and those affecting the manner in which forests modify snow deposition on the ground. These two groups can be treated effectively as abiotic and biotic components of the larger system. The abiotic components involve synoptic weather patterns, cooling of air masses, initiation of snow crystals, and the influence of wind and slope upon snow delivery. These components follow physical relationships over which man has little influence (cloud seeding is an exception). They are treated first (Section II) and indicate the boundary conditions influencing snow deposition in southwestern B.C. The influence of forest cover on snow deposition is treated in Section III. Although physical relationships are still fundamental, man can exploit these by altering the forest structure. Section IV summarizes the implications derived from the material reviewed in II and III.

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II. ABIOTIC FEATURES AFFECTING SNOW DELIVERY

2. AN OVERVIEW AND CONCEPTUAL FRAMEWORK

This chapter has two objectives: i) to present the broad conceptual framework which has determined the structure of Chs. 3 and 4; and ii) to state briefly the organization and contents of all chapters following. The latter objective is meant as a convenience to the reader. Some sections are long and all are interrelated. A brief "reader's guide" should help a reader both to understand our approach in pursuing the objectives of this report (p. 7 and 8), and to find material relevant to a particular question more quickly.

2.1 Conceptual Framework - Abiotic Processes

Part II deals solely with delivery of snow to a surface. We segregate two broad groups of processes; those governing the production of snow (Ch. 3), and those governing the delivery of snow (Ch. 4). At the detailed level of cloud microphysics, the physical processes which determine the probability of snow production and the kinds of snow produced should apply universally. The processes yield different outcomes primarily as functions of the temperature and moisture of the air mass in which they occur. The temperature and moisture of the air mass are themselves functions of the source of the mass and lifting processes which produce cooling (Fig. 2.1).

The empirical observations summarized are always the results of simultaneously occurring processes. Any attempt to segregate processes is thus somewhat artificial. However, failure to segregate processes severely limits the ability to extrapolate (Ch. 1.1). Our approach has been to document both empirical observations for southwestern British Columbia and apparent generalities extracted from wide-ranging literature. We then evaluate the apparent consistency of literature-based relationships with more local, empirical observations. Where consistency is evident or changes predictable we have approximated objectives 3), 5), and 6) (p. 8).

The conceptual framework as presented here is brief; it is enlarged and presented in more detail in the relevant sections of Chapters 3 and 4. The framework begins with synoptic features of coastal snow storms (Ch. 3.1). These features are the result of the origin of air masses involved and their trajectory towards the coast. The origin and trajectory determine the temperature, moisture content, and concentrations of cloud condensation nuclei. The mixing of different air masses determines the kind of storm that will occur (Fig. 2.1). The air mass that results represents the raw material upon which lifting processes act and within which microphysical processes occur. Air masses and their trajectories are thus the first determinants of the probability of snow. Because air masses arriving at the coast

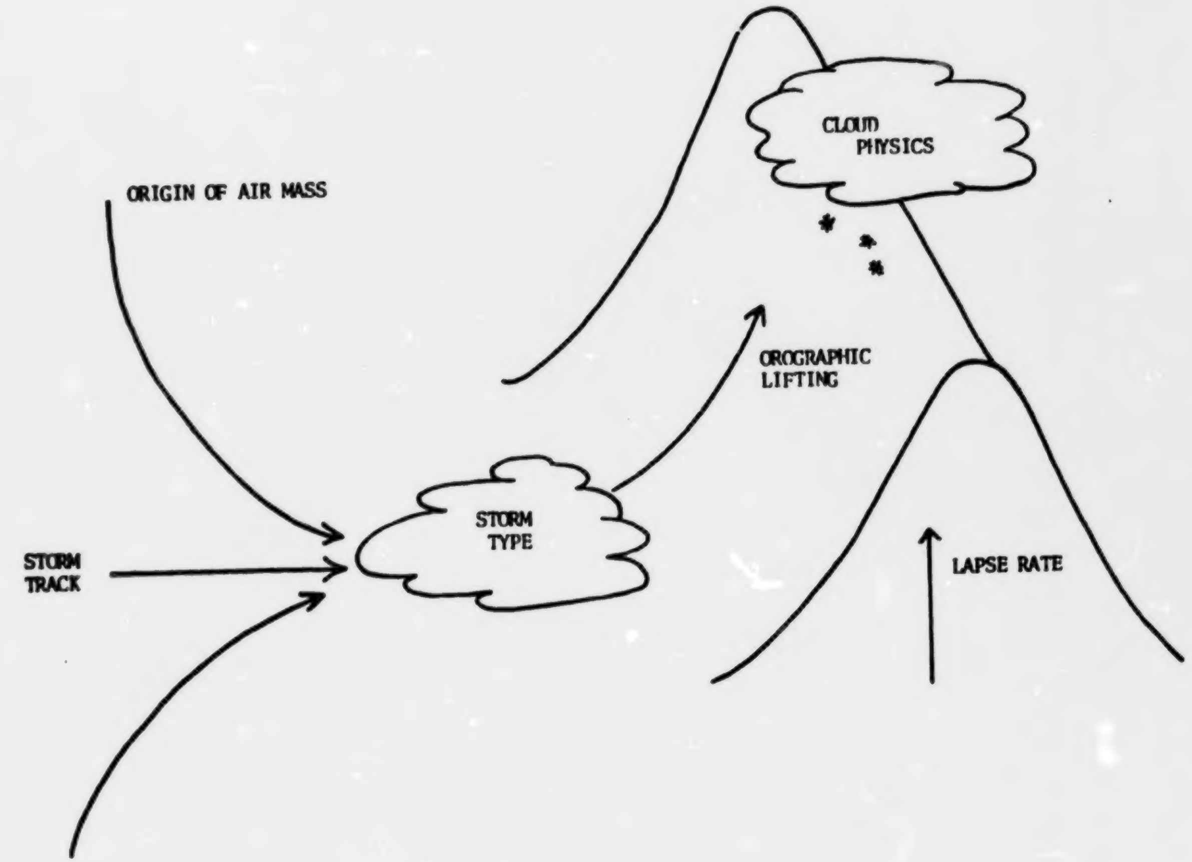

ORIGIN OF AIR MASS
STORM TRACK
STORM TYPE
OROGRAPHIC LIFTING
CLOUD PHYSICS
LAPSE RATE

Figure 2.1. Schematic representation of major factors creating snow.

are variable (Fig. 3.4) and follow a variety of storm tracks (Fig. 3.6), only probabilities can be extracted (e.g., Tables 3.1 and 3.2).

Few air masses, or mixtures of air masses, arriving at the coast have the appropriate temperature and moisture contents to generate snowfall at sea level. Moisture content is rarely limiting to snowfall (Fig. 3.5) but some lifting and cooling must occur (Fig. 3.1) before the air is saturated. Review of lifting processes and existing data indicates that orographic lifting is favoured by coastal conditions and dominates all lifting processes, and that the empirical lapse rate is near-neutral (about $7^{\circ}\mathrm{C}\cdot\mathrm{km}^{-1}$). The model can thus be simplified to consider orographic effects on precipitation and the proportion of precipitation that falls as snow (Ch. 3.2). These relationships are empirical, the result of particular air masses subject to orographic lifting (e.g., Table 3.4).

Whether snow actually forms and falls or not, is dependent on the processes of cloud physics and wind speed (Fig. 2.1). Lifting acting on air masses simply determines the environment in which these latter processes interact. Our approach is to review the fundamental nature of the processes, then to evaluate their consistency with empirical observations (Ch. 3.3). The processes involved are presented in more detail in Figure 3.12. Sufficient data exist to document that most processes in southwestern British Columbia act as predicted from general relationships. For example, the kinds of ice crystals formed are those that are predicted from basic physical relationships. However, we were unable to evaluate unequivocably any potential limitation imposed by concentrations of cloud condensation nuclei.

Chapter 3 does no more than examine the processes creating snow with particular reference to southwestern British Columbia. It provides a more detailed conceptual framework of these processes than is presented here, documents the relationships most likely to be important in British Columbia (objective 5), and evaluates the likelihood that relationships developed elsewhere are applicable to coastal British Columbia (objective 6). The ability to associate the few scattered observations from southwestern British Columbia with potentially general relationships provides more confidence to extrapolation and allows us to work towards the boundary conditions of objective 3. Other chapters utilize an analogous approach with their subject matter.

Chapter 4 begins with the snow created (Ch. 3) and considers how slope and wind would affect snow delivery to a surface. It is more speculative than Chapter 3 because we could find no theoretical framework to adapt and have devised our own. The intent is that upon completion of Chapter 4, important abiotic processes have been addressed and the 'snow can meet the trees'. The relevant processes occur at two broad levels: i) the macroscale of mountain slopes; and ii) the interactions of wind velocity with surface roughness and of snow particles with wind velocity (Fig. 2.2). Because we are concerned ultimately with processes within a forest stand,

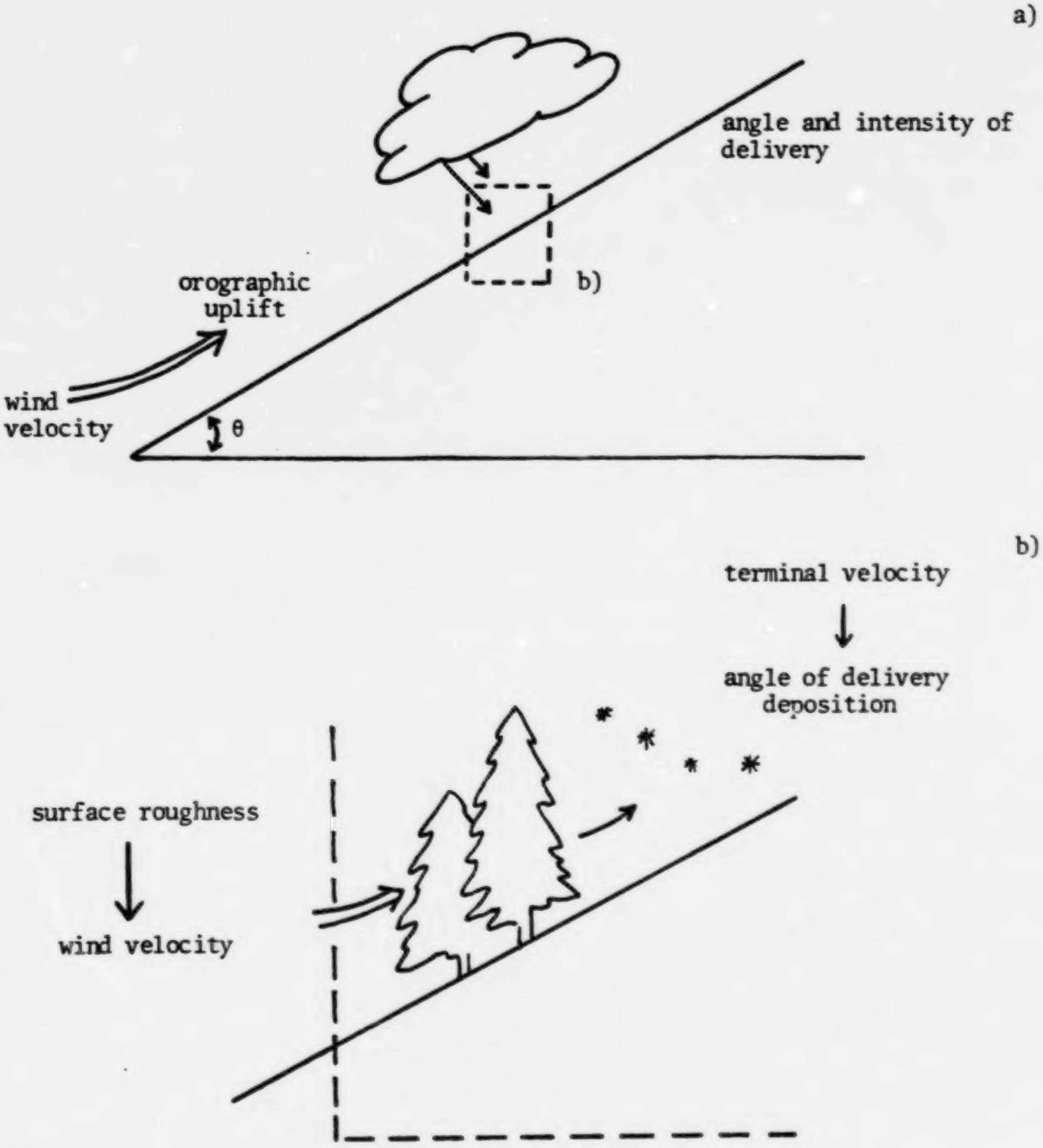

a)
angle and intensity of delivery
b)
orographic uplift
wind velocity
θ
b)
terminal velocity
↓
angle of delivery deposition
surface roughness
↓
wind velocity

Figure 2.2 Schematic representation of major factors affecting snow delivery to a surface.

Chapter 4 serves primarily to deal with the broad form of the processes. Specifics are addressed in Chapters 7 and 8 when characteristics of individual trees and stands are introduced.

Were snow to fall vertically, snow on a slope would be some function of the cosine of slope angle. Snow seldom falls vertically, nor is the snow-forming surface necessarily parallel to the receptor surface. To a considerable extent the orientation of snow-forming and receptor surfaces, and the vector of snowfall, are functions of wind speed. Wind speed and slope thus interact. The interactions are noted in Ch. 4.1, but predictions are not tested until Chapter 7. As discussed in Chapter 4, the interactions are for large, planar surfaces such as mountain slopes (Figs. 4.1 and 4.2); they are tested for the smaller surfaces of a tree crown. Between these two scales, friction or surface roughness alters wind speed. Roughness coefficients of forests are relatively large, varying from 30 to 290 cm. Forest cover thus both reduces wind velocity and generates turbulent flow at low wind velocities. Both effects reduce the likelihood that snow will be carried in the air column rather than delivered to some intercepting surface. Whether the snow is deposited from the air stream is a function not only of the velocity and turbulence of the stream, but also of the terminal velocity of the snow particle (Fig. 2.2). Terminal velocity is itself a function of bulk density and drag coefficients. Given the nature of snow particles frequently formed in coastal British Columbia (Ch. 3.3), the terminal velocities are relatively high (Fig. 4.8). The review of processes in Chapter 4 suggests relatively dense new-fallen snow, relatively small effects of wind during snow delivery, and pronounced effects of forest cover on snow deposition from the air stream. Transport of deposited snow is not addressed in Chapter 4, but is treated in the discussion of opening size (Ch. 8.3). The conceptual framework appropriate to interception in forests is presented in Chapter 6.

2.2 The Pieces and the Whole - A Reader's Guide

All chapters exploit material presented in other chapters. The overview presented here is meant to clarify how these separate pieces relate to the whole. It indicates briefly the philosophy that led to the report's structure, and serves as an adjunct to the table of contents. The report begins with physical or abiotic processes that would occur whether or not a forest stand were present. It then relates these processes to individual trees and lastly to whole stands. At each stage more variables are aggregated and predictability decreases. The final chapter summarizes the management implications evident in the review. To reduce repetition, relevant points are treated briefly.

Chapter 3: describes the general processes creating snow, documents specific relationships dominant in southwestern British Columbia and potential factors limiting to snowfall, provides initial estimates of governing relationships, and describes the kinds of snow that occur.

Chapter 4: provides an overview of the manner in which slope and wind potentially affect snow delivery (including factors modifying wind), and documents empirical terminal velocities of snow particles common to coastal mountains.

Chapter 5: because they discuss general relationships as well as those dominant in southwestern British Columbia, Chapters 3 and 4 are lengthy. Chapter 5 summarizes the key findings developed from the review of abiotic relationships. It serves to set the stage for discussion of interception processes. The summary is in point form. Readers wishing only the essence of Chapters 3 and 4, should read Chapter 5.

Chapter 6: presents the definitions used and the conceptual framework underlaying the structure of Chapters 7 and 8. It is essential reading if the analyses of Chapter 7 are to be appreciated fully. Interception is first treated specifically in Chapter 6.

Chapter 7: introduces two physical processes not evaluated previously (adhesion and cohesion), then attempts to evaluate individual factors affecting interception by single trees. Abiotic factors (wind, temperature, storm size) are first evaluated in terms of predictions from Chapters 3 and 4, then related to individual attributes of the tree crowns. Attributes of whole crowns are analyzed in terms of their ability to predict interception; their utility for prediction is ranked and explained in terms of the conceptual model of Chapter 6 and the details of Chapters 3 and 4. This chapter provides the most detailed description of interception and forms much of the basis for Chapter 9.

Chapter 8: treats processes at the level of individual stands and deals with much broader measurements which serve as surrogates for specific variables treated in Chapter 7. Aggregation of processes is continued. Four broad factors are evaluated (storm size, elevation, crown closure, and size of forest openings); only storm size can be compared to predictions for individual trees. The treatment of wind, first introduced in Chapter 4, is refined to incorporate stand characteristics (permeability and opening size). The analyses based on data for individual storms are then evaluated against the still broader measurements of snowpack.

Chapter 9: summarizes the implications that can be drawn from the preceding analyses in terms of their relevance to prescriptions for developing winter range in managed stands. Because the treatment is in point form without continuous qualification, it is potentially misleading without reference to the other chapters from which implications are summarized (brevity may appear as confidence). Suggestions for subsequent research are presented.

3. PROCESSES CREATING SNOW

Snowfall is a particular type of precipitation - it shares similarities with but also differs from rainfall. In general, the occurrence of precipitation is determined by the availability of atmospheric moisture and the presence of mechanisms which can convert the moisture into precipitation. The primary mechanism is cooling resulting from the vertical motion of the air, until the air can contain no more moisture and condensation occurs around small particles in the air mass. For snow to occur the temperature must be at or below freezing at some height in the air mass (snowfall can occur when ground surface temperatures are above freezing). Snow can be broadly and loosely defined as particles of ice formed in a cloud which have grown large enough to fall with a measurable velocity and reach the ground.

The primary process by which moist air cools and produces precipitation is adiabatic; there is no net gain or loss of heat in the entire system. Adiabatic cooling occurs as air masses lift vertically and gain altitude. The kinetic energy of molecular motion (indicated by temperature) is utilized to expand the air parcel against 'surrounding' pressure. Thus, as elevation increases both air pressure and air temperature (molecular motion) decrease. For a more complex treatment see Pruppacher and Klett (1978: 350-357).

The rate at which air temperature decreases with increasing elevation is termed the adiabatic lapse rate. In dry air, the rate is about $10.0°C\cdot km^{-1}$ ($9.76°C$ if air were an ideal gas). If air contains moisture, a temperature is reached at which condensation occurs. Condensation releases the latent heat of vapourization and slows the rate of temperature decline. Moister air therefore has a slower lapse rate. Warm air also has a slower lapse rate than cool saturated air because it can contain more moisture. Saturated adiabatic (or pseudo-adiabatic) lapse rates for lifted air parcels in cold air are $\geq 9°C\cdot km^{-1}$ but only about $4°C\cdot km^{-1}$ in warm dry air.

As air cools it can retain less moisture. The amount of cooling to produce supersaturation depends on moisture content and initial temperature (thus, initial relative humidity). For example, a typical air mass from the Pacific containing $9.4\ g\cdot m^{-3}$ water vapour need only be cooled below 10°C before becoming supersaturated (Fig. 3.1).

Fitzharris (1975) determined the lapse rate during winter storms on Mt. Seymour B.C. to be $7°C\cdot km^{-1}$ which is probably a good first approximation to lower elevation lapse rates in coastal B.C. during the winter. During coastal winter storms there is often a frontal or marine inversion based from 1000 to 2000 m above the surface (but that is well above winter range). A lapse rate of $0.7°C\cdot 100\text{-}m^{-1}$ falls within the range of pseudo-adiabatic lapse rates given in the Smithsonian Meteorological Tables for the conditions of most storms at about 950 m (i.e., pressure 950 to 850 mb, temperatures -10 to 15°C). Given the empirical lapse rate of $7.0°C\cdot km^{-1}$ we can

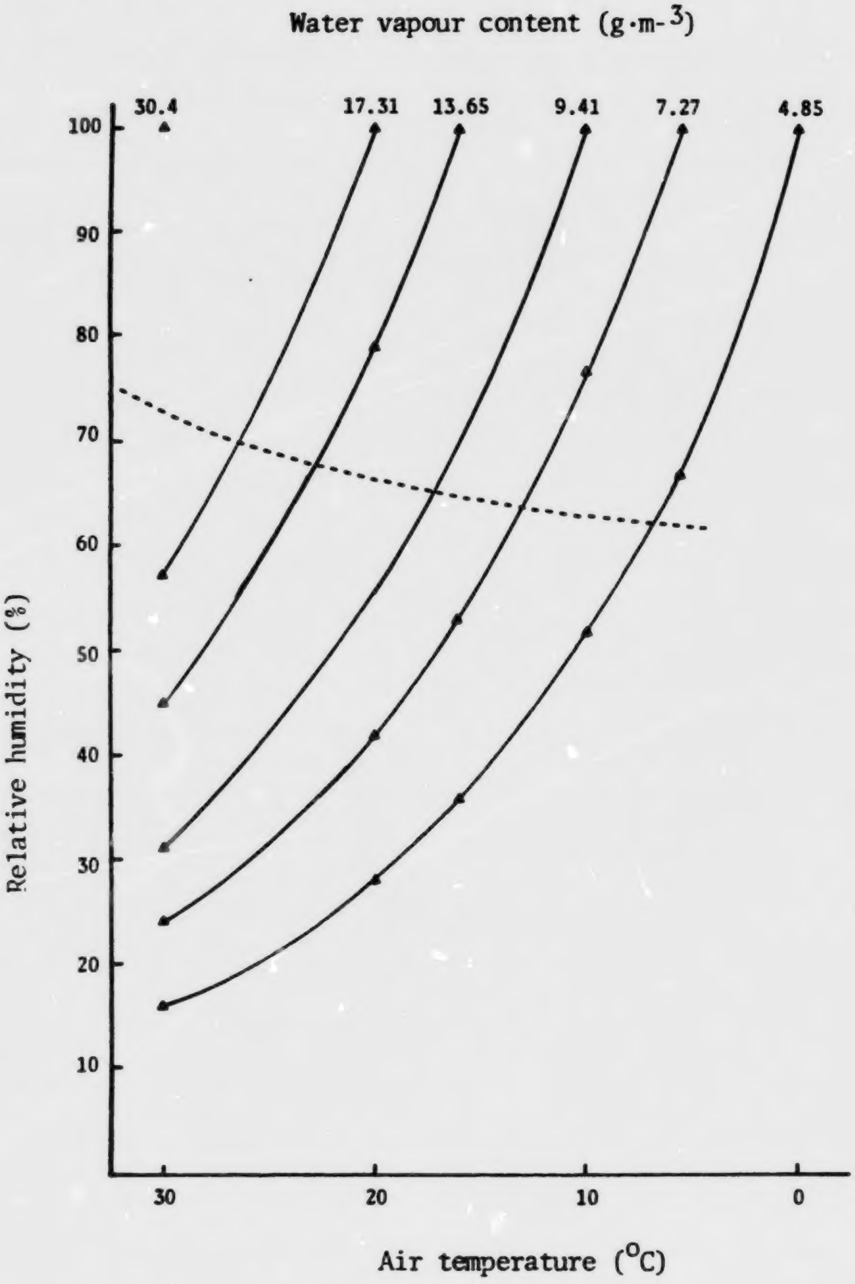
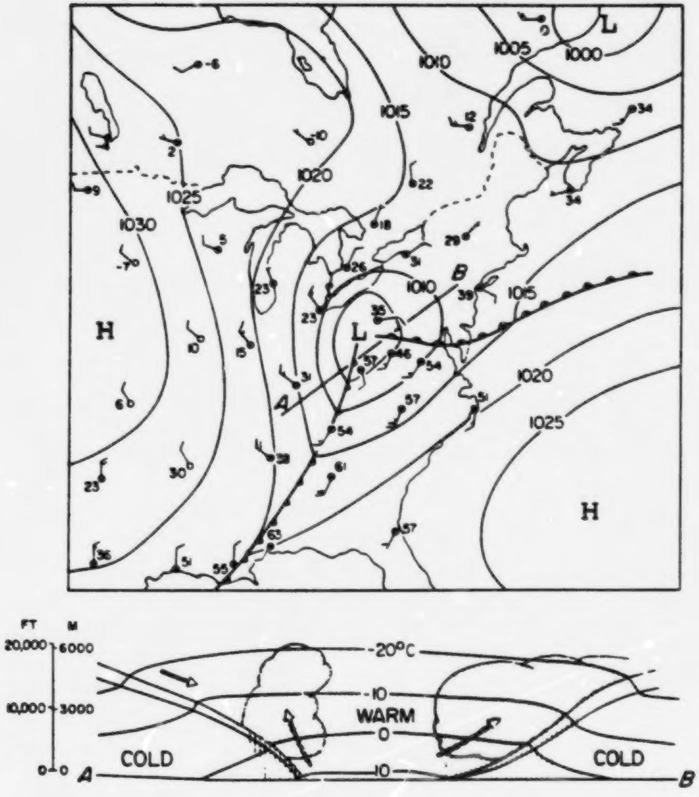


Figure 3.1 Capacity of the atmosphere to retain moisture as a function of temperature (data for solid lines from McKay 1970). At any point on the dashed line a cooling of 7°C would result in 100% relative humidity; given the empirical winter lapse rate, that cooling would result from a 1000 m elevation gain.

compute the effect of a 1000 m elevation gain on moisture-holding capacity of the air in southwestern British Columbia (dashed line of Fig. 3.1). Most incoming warm air masses containing more than 5 $g \cdot m^{-3}$ water vapour would become supersaturated if they rose more than 1000 m.

The lapse rate applies when adiabatic cooling is invoked by air masses rising. Vertical motion of air masses occurs in four major ways. The first, 'frontal lifting', is associated with low pressure areas and frontal systems. When warm and cold air masses meet, the warm air rides up over the cold air, decreasing in temperature as it does so. The resulting frontal zone is usually 100-200 km wide and is characterized by precipitation if the warm air is sufficiently moist (Fig. 3.2). The second means is termed 'orographic lifting' and occurs as an air mass is thrust against a vertical topographic barrier, typically a mountain range. The air is forced to rise and therefore cool adiabatically. Consequently, the windward sides of mountain ranges are characterized by high precipitation regimes (Fig. 3.3) as can be seen from any continental precipitation map. The third means is 'convective lifting' resulting from warming of surface air by contact with warm ground or, more commonly, warm water. Convective currents typically result in piled-up, cumuliform clouds and thunderstorms or squally, showery snowfall (Fig. 3.3). Convective lifting can be intensified by orographic lifting. 'Horizontal convergence', the fourth process generating vertical motion, may occur as a result of a


L
1000
1005
1010
1015
1020
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H
L
1010
1015
1020
1025
H
A
B
FT
M
20,000
6000
10,000
3000
0
0
-20°C
-10
WARM
0
10
COLD
COLD
A
B

Figure 3.2 Typical low pressure area showing a cross-section (A-B) of warm and cold fronts (from Richards 1973: 11).

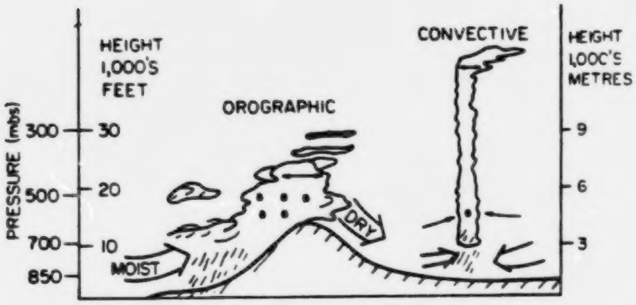

HEIGHT 1,000'S FEET
CONVECTIVE
HEIGHT 1,000'S METRES
OROGRAPHIC
PRESSURE (mbs)
300
500
700
850
30
20
10
9
6
3
MOIST
DRY

Figure 3.3 Schematic representation of orographic and convective lifting (from Richards 1973: 11).

wind field directing the flow of air into a particular area; e.g., a surface low pressure area causing air to lift. Horizontal convergence in the lower atmosphere is normally associated with frontal systems which lie in low pressure troughs. Divergence aloft generates vertical motion.

Even when the air mass is lifted, condensation does not occur easily in a pure environment. It occurs much more readily when the air mass contains an ample supply of very small particles (aerosols) of a hygroscopic nature which act as nuclei for the condensing vapour (here termed CCN, cloud condensation nuclei). Pure water vapour at room temperature will not condense until it is super-saturated to four times the vapour pressure needed to condense water vapour when condensation nuclei are readily available. In coastal British Columbia chloride nuclei, formed from bursting salt water bubbles caused by ocean waves, could provide abundant CCN.

For snow to fall rather than rain, the temperature at condensation must be below freezing. The microphysics of clouds and precipitation determine the amount of water condensed and whether that water, in whatever form, reaches the ground. Ludlam (1955, 1956), estimated that the actual formation of the precipitation particle (rain drop or ice crystal) requires 0.5 to 1.0 h depending on the process involved.

Summarizing at the broadest level, we note four major processes involved in creating snowfall:

1) atmospheric lifting of the air mass,

2) presence of sufficient water vapour in the lifted air mass,

3) available nuclei for condensation, and

4) temperature near or below freezing.

For coastal British Columbia we can concentrate on orographic lifting because it almost always is involved. The mountainous nature of the coast means that orographic lifting augments amounts of precipitation generated by frontal or other lifting (Ch. 3.1). On eastern Vancouver Island, convection involving over-water trajectories in winter may be the primary cause of lifting and precipitation, but orographic effects will again contribute. Conditions favouring orographic precipitation include: i) strong winds perpendicular to mountain ridges, although stable laminar air flows over a barrier can occur; ii) an air mass which is moist in depth, preferably containing existing cloud layers; and iii) an air mass with a near neutral lapse rate (without markedly stable layers or inversions). Such conditions appear to prevail on B.C.'s coast (Ch. 3.1).

Several theoretical models for estimating orographic precipitation have been proposed (Sawyer 1956, U.S.W.B. 1961 and 1966, Walker 1961, Elliott and Shaffer 1962, Danard 1971, Fitzharris 1975, Hetherington 1976). However, some aspects of the theory of orographic precipitation remain little understood. Much of the theory deals with ideal situations (e.g., an infinitely long, smooth mountain ridge), and some theories incorporate terms which cannot be measured with precision (e.g., precipitation efficiency, terminal velocity of precipitation particles). All theoretical models treating orographic precipitation on a daily basis do have one common feature: they all incorporate meteorological aspects at three different scales. First, are the large-scale, synoptic factors which determine the characteristics of the air masses crossing the mountains (e.g., wind speed, direction, stability, and humidity). Second, are the dynamics of air motion over a mountain ridge (through what height and layers are air masses lifted). Third, are the microphysics of clouds and precipitation formation, particularly the proportions of condensed vapour reaching the ground or evaporating on the far side of the ridge. We treat each scale separately (Ch. 3.1 to 3.3).

Walker (1961) and Danard (1971) applied their theories to various areas of British Columbia, but both sought to estimate seasonal or annual precipitation for mountain areas which were topographically smoothed over 200 km^2 or greater. Woo (1972) derived empirical equations describing daily rainfalls at the UBC Research Forest; Hetherington (1976) treated the orographic component. Our concern here is the prediction of orographic precipitation after individual storms for a mountain area of about 0.5 to 1.0 km² and is thus similar to the approach of Fitzharris (1975). The following three sections examine the four major processes involved in the formation of snow at the three scales necessary in any treatment of orographic precipitation.

3.1 Synoptic Features of Coastal Snow Storms

Air masses are classified on the basis of two primary factors. The first is temperature (giving Arctic, Polar, and Tropical air); the second is surface type in their region of origin (giving maritime and continental categories).

The source and trajectory of air masses determine their temperature and humidity upon arrival at the coast. Broad patterns of air movement thus determine the raw material upon which orographic effects and adiabatic lapse rate will operate.

During winter, British Columbia can experience a wide variety of air masses (Fig. 3.4). The continental Arctic air mass (cA of Fig. 3.4) originates much farther north than either cP, mP, or MT, and consequently is cold and dry. Generally it remains east of the Rockies as it moves south, but occasionally breaks out into British Columbia. Kendrew and Kerr (1955) suggested that, on average, such invasions of continental Arctic air only reach the coast once every two years. However, Fitzharris (1975) recorded 8 such outbreaks in the winter of 1970-71 and one in the winter of 1969-1970.

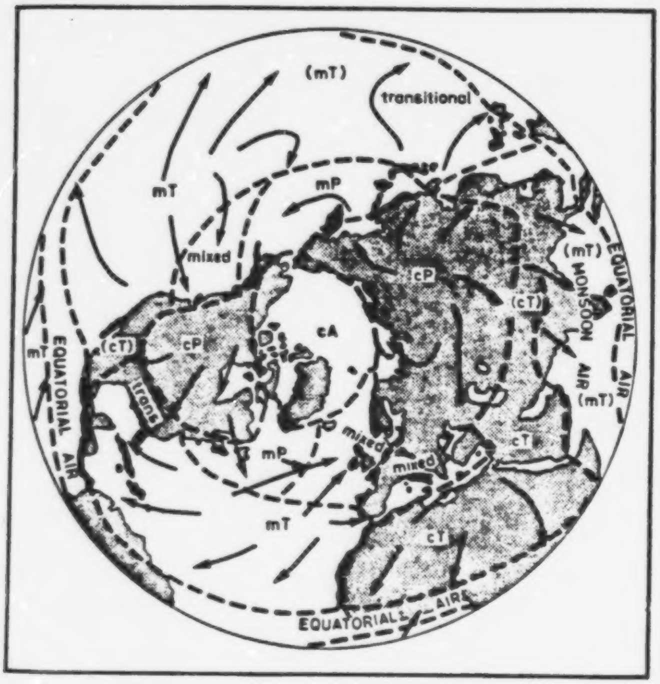

(mT)
transitional
mT
mP
mixed
(mT)
EQUATORIAL
MONSOON
AIR
(cT)
cP
cA
(cT)
cP
mT
EQUATORIAL AIR
trans
mP
mixed
mixed
cT
mT
cT
EQUATORIAL AIR

Figure 3.4 Air masses in winter (from Barry and Chorley 1976: 182). Legend: A = Arctic, P = Polar, T = Tropical, c = continental, m = maritime.

Fitzharris apparently did not distinguish between cA and cP air masses. However, differences between cP and cA air masses are limited mainly to the middle and upper troposphere, where temperatures are lower in the cA air (Godson 1950). The effect of either air mass upon precipitation at the coast would be similar.

Kendrew and Kerr (1955) recognized three types of maritime air masses affecting British Columbia. Cold maritime air originates in the cold, dry interior of central Asia and is warmed and humidified as it moves across large expanses of ocean. The degree of modification depends on the amount of time spent over the water and therefore on the trajectory. A short passage results in cold, fairly dry air termed maritime Arctic air. Longer passage results in a much warmer and more humid air mass termed maritime Polar (mP of Fig. 3.4). At times the two cold, maritime masses are so different in character that a frontal discontinuity separates them, resulting in a "maritime Arctic front". The dominant mechanism generating precipitation at the B.C. coast is the passage of frontal disturbances on the maritime front (between mA and mP air masses). The maritime Tropical (mT) air masses originating in the sub-Tropical Pacific, are of less direct importance on the British Columbian coast. However, maritime Tropical air often can be found at high tropospheric levels in the systems involving Arctic and Polar air. In each case the amounts of precipitation which fall at the surface are strongly modified by orographic effects.

Tropical air masses do have an important indirect effect. The confluence of Arctic and Tropical air masses creates a zone of high cyclonic or frontal activity in the south coastal area and south of the Aleutian Islands. From October to mid-May, weather is dominated by the "Aleutian Low", a semi-permanent pressure system located south and east of the Aleutians. This system spawns series or interconnected families of four or five pressure cells which may affect the coast for two weeks or more. The analyses of Klein (1957), and Reitan (1974) indicated that coastal B.C. is the preferred track for cyclonic disturbances. Maunder (1968) stated that during January of both 1964 and 1965 lows occurred on 72% of the days.

One result of these broad patterns is that the North Pacific is a major source region for moisture which is carried eastward by the prevailing westerly circulation (Schemenauer et al. 1981). In winter, the maximum moisture transport across the Pacific probably occurs between latitudes of 50 and 55°, with the mean flow decreasing almost to zero between 30 and 35° N (Rasmusson 1967). When considering the distribution of atmospheric moisture associated with transport patterns, it is helpful to use the concept of 'precipitable water' (the depth to which liquid water would stand if all water vapour were condensed to a vertical atmospheric column of uniform cross section). Hay (1970) computed the distribution of mean precipitable water for Canada; the January distribution is illustrated (Fig. 3.5). The largest values (10 to 12 mm)

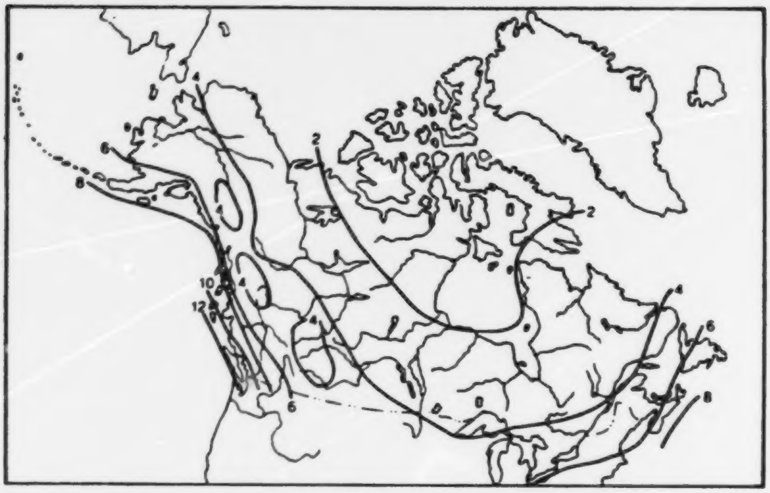

2
2
4
4
4
4
6
6
6
8
8
10
12

Figure 3.5 Distribution of the mean depth of precipitable water (mm) over Canada in January (from Hay 1970).

occur along the southern British Columbia coast and are a function of relatively warm, moist air from the Pacific.

The broad pattern of Figure 3.5 is supported by specific observations. Fitzharris (1975) reported that most snow storms on Mt. Seymour were frontal (cyclonic) in nature and dominated by cold maritime air masses (Table 3.1). Storms involving maritime Tropical air were less frequent but caused rain or rain and snow mixtures at all elevations up to 1260 m (presumably because they were warm and moisture laden). Colder air usually invaded behind fronts and lowered the freezing levels. Thus, when ambient temperatures were warm, most snow fell post-frontally. In general, however, most snow was produced pre-frontally. The heaviest snowfalls were associated with the relatively rare outbreaks of continental Arctic air.

Fitzharris (1975) defined 8 generalized storm tracks (Fig. 3.6). Storms following tracks A and B generally did not have low enough freezing levels to produce snow at elevations below 1260 m. Storms following tracks E, G, and H had the greatest access to cold air sources and commonly produced snow with a large orographic component. Storms along tracks E and C were strongly influenced by ocean temperature and therefore produced more snow at lower elevations in April than in November when they contained less moisture. Storms on track G produced snow most frequently (93% of the time), probably because the air mass was initially cold and acquired moisture to near saturation level during the long maritime trajectory

Table 3.1 Percent frequency of storm types for all storms and for snow storms on Mt. Seymour during the winters 1969-70 and 1970-71. Storm types were identified from surface synoptic charts. There were 73 storms in the winter 1969-70 and 74 storms in the winter 1970-71 (data of Fitzharris 1975).

Storm Type	All Storms		Snow Storms	
	1969-70	1970-71	1969-70	1970-71
A. Frontal	77	89	45	65
1) Maritime cold fronts	12	19	7	15
2) Maritime fronts with a warm sector	15	18	7	10
3) Maritime occluded fronts	48	39	30	27
4) Arctic fronts associated with maritime front	2	8	1	8
5) Arctic fronts not associated with maritime fronts, but with moist Pacific airflows	0	5	0	5
B. Non-frontal	23	11	15	8
6) Cold lows	6	0	1	0
7) Other lows with no front, or front well to S	5	4	6	3
8) Troughs (usually associated with W or NW airflow	2	5	1	4
9) SW airstreams	3	1	3	0
10) W airstreams	0	1	0	1
11) NW airstreams	4	0	3	0
12) Light precipitation associated with high pressure cells	3	0	1	0
TOTALS FOR WINTER	100%	100%	61%	73%

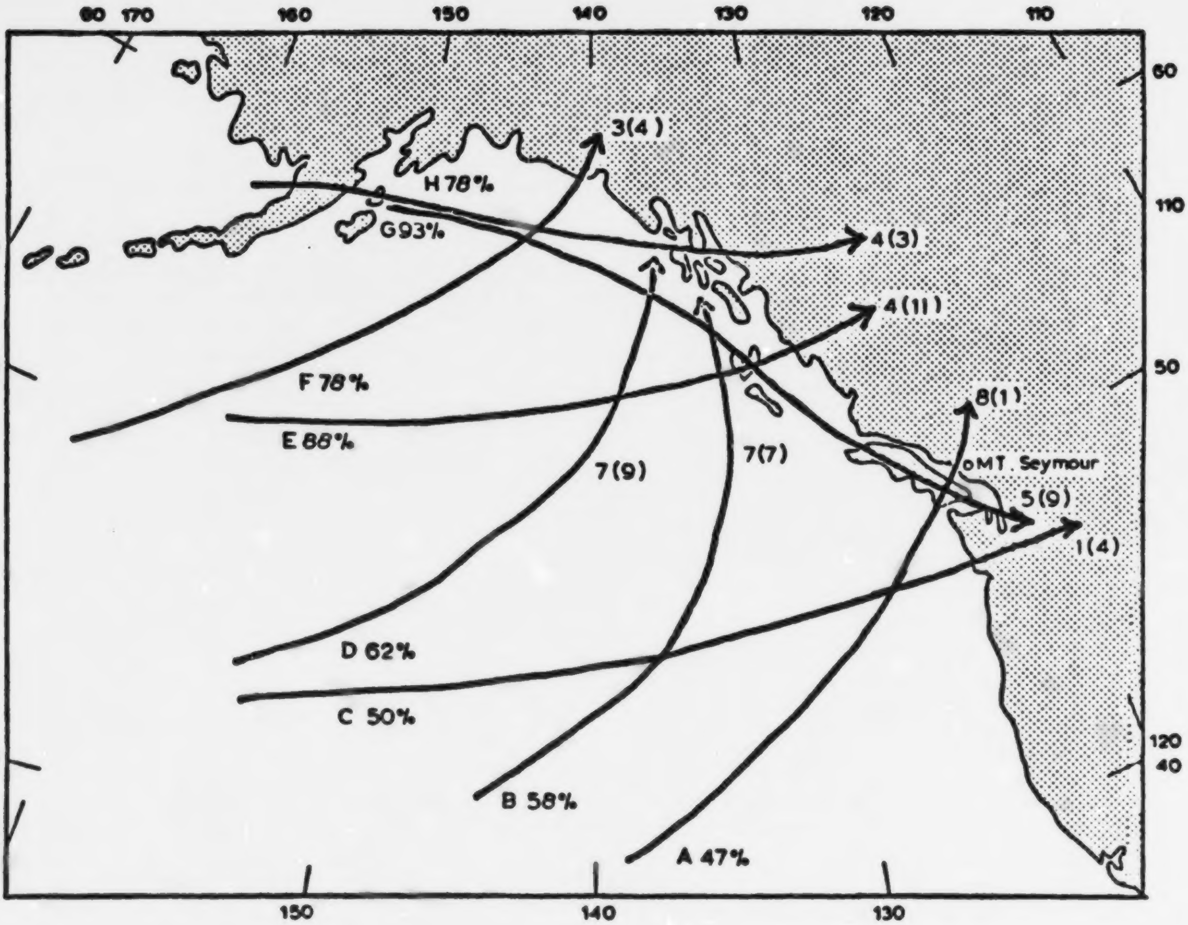

80
170
160
150
140
130
120
110
60
110
50
120
40
170
50
40
160
150
140
130
3(4)
H 78%
G 93%
4(3)
4(11)
F 78%
E 88%
8(1)
7(9)
7(7)
MT. Seymour
5(9)
1(4)
D 62%
C 50%
B 58%
A 47%

Figure 3.6 Generalized tracks of storms (letters) off British Columbia coast during the winters of 1969-70 and 1970-71. Percentages are percent of storms following each track that produced snow on Mt. Seymour. Numbers at the ends of the arrows indicate the frequency of snow storms with those for 1970-71 in brackets (from Fitzharris 1975: 201).

(Fig. 3.6). Any additional cooling usually resulted in snow. Storms following any track can produce heavy snowfalls when associated with Arctic air outbreaks.

Storms distributed snow differentially on the mountains depending upon the storm type. On Mt. Seymour, occluded fronts and cold fronts made the most significant contribution. An occlusion occurs when fresh, cold air in the rear overtakes the modified cold air at the front of a depression. The warmer air is pinched off and lifted from the surface. Only at high elevations (>800 m) did warm fronts generally produce any snow at all. At middle elevations, occluded fronts were important as well as Arctic air outbreaks and cold fronts. At low elevations (<400 m) Arctic air tended to dominate snowfall. Cold fronts and troughs also made a contribution but occluded fronts were important only when they had access to cold air.

At high elevations, the greatest snowfall per storm was associated with southwest airstreams, Arctic air outbreaks and cold fronts. At middle elevations, storms associated with Arctic air, cold fronts, and a few non-frontal types produced the heaviest snowfall. At low elevations only Arctic air outbreaks produced large amounts of snow. These empirical observations broadly fit expectations derived from the characteristics of specific air masses. For example, only the Arctic air masses would likely be sufficiently cold to produce snow at low elevations.

Table 3.1 presents Fitzharris' (1975) classification of storm types on Mt. Seymour during the winters of 1969-1970 and 1970-1971. Maritime occluded fronts were by far the most common storm type (48% in 1969-70 and 39% in 1970-71). It is possible to calculate conditional probabilities of snowfall given a particular storm type from the data provided by Fitzharris. Such probabilities and other aggregated statistics are presented in Table 3.2. During the winters of 1969-70 and 1970-71 maritime fronts constituted 75% of all storms (n = 147) and 76% of the storms producing snowfall (n = 98). Conditional probabilities of snowfall given a particular storm type were generally greater in non-frontal storms, but such storms occured with a frequency of less than 20% (Table 3.2). On Mt. Seymour the joint probability of a particular storm type occurring and producing snow was three times greater for maritime occluded fronts than for any other type (P = 0.30, Table 3.2).

If the data for Mt. Seymour are generally representative of southern coastal conditions we can characterize the broad scale synoptic features necessary for any treatment of orographic effects. Frontal systems represented 83% of the storms and 84% of those producing snow. Only 8% of the fronts involved drier Arctic air masses (Table 3.2). The arriving air was moist in depth and sufficiently unstable to discourage the creation of inversions at these elevations. The adiabatic lapse rate was thus lower than that of dry air and showed few marked discontinuities (near neutral). We should expect a strong orographic component. Marine and frontal inversions do

Table 3.2 Probability of occurrence of different storm types and conditional probabilities of snow given a specific storm type. Data for Mt. Seymour during winters 1969-70 and 1970-71 (calculated from Fitzharris 1975: Table 8.1).

	Number of storms	Probability[1] of occurrence	Conditional probability of snow given specific storm type			Probability of snow and storm[2]
			1969-70[2]	1970-71[3]	both winters	
Frontal	122	0.83	0.59	0.74	0.67	0.56
1) Maritime cold	23	0.16	0.56	0.71	0.65	0.10
2) Maritime warm	24	0.16	0.46	0.77	0.63	0.10
3) Maritime occluded	64	0.43	0.63	0.76	0.69	0.30
4) Arctic front with maritime front	7	0.05	1.00	0.67	0.71	0.04
5) Arctic front with moist Pacific flow	4	0.03	-	0.75	0.75	0.02
Non-frontal	25	0.17	0.65	0.75	0.74	0.13
6) Cold low	5	0.03	0.20	-	0.20	0.006
7) Other lows	7	0.05	1.00	0.67	0.86	0.04
8) Trough	4	0.03	1.00	1.00	1.00	0.03
9) SW airstream	3	0.02	1.00	0.00	0.67	0.01
10) W airstream	1	0.01	-	0.00	1.00	0.01
11) NW airstream	3	0.02	0.67	-	0.67	0.01
12) Light precipitation	2	0.01	0.5	-	0.50	0.005

1/ out of 147 storms
2/ out of 73 storms
3/ out of 74 storms

occur, but their base is usually above the elevations of winter range.

3.2 Elevation and Snowfall

Snow on the ground increases with increasing elevation for three reasons: i) increasing elevation causes increasing precipitation (orographic effect); ii) in mild, coastal areas, low elevations are commonly above freezing and precipitation in the form of snow is more common at higher elevations; and iii) lower temperatures at high elevations retard melt. The melt and delivery processes are thus inextricably linked. Most data available refer to occasional snowpack measurements suitable for hydrological research or forecasting, and therefore fail to differentiate between delivery and melt phases. Here the few data relevant to the effect of elevation on snow delivery are treated.

We adopt the perspective that snow delivery varies with elevation due to: i) the orographic effect on precipitation; and ii) the temperature-related proportion of precipitation falling as snow.

3.2.1 Orographic Effects on Precipitation

Relatively little is known about air flow over mountains. One important feature is that at some height "h", the upward motion of the air is reversed (Fig. 3.7). For effective

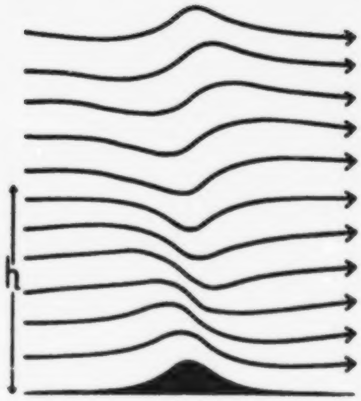


Figure 3.7 Schematic representation of streamlines in air passing over a ridge. The upward motion of air immediately above the ridge is reversed at some level (h) and replaced by downward motion (from Sawyer 1956: 376).

orographic precipitation to occur, h must be large enough for sufficient adiabatic cooling to have occurred. Sawyer (1956) suggested that conditions favouring a large value of h occur with a fast-moving air mass containing existing cloud and a lapse rate near saturated adiabatic. Clearly, larger mountains should also increase h.

The speed at which ice crystals form is also important. If the process requires large amounts of time, the supersaturated air mass may have moved over the mountain only to descend and re-evaporate the particles or dump the snow on a further mountain range (Fig. 3.7). Available data (Ch. 3.3) suggest that there is usually sufficient time for particle formation in south coastal British Columbia.

Mathematically, the precipitation rate is predominantly related to terrain slope and wind flow rather than to elevation. If the air is saturated, the rate at which precipitation is produced is directly proportional to the ascent rate of the air mass which is a product of wind speed and slope angle (Eq. 4.4). Rhea and Grant (1974) provided a general mathematical expression for orographic snowfall in mountains. They included the effects of large-scale vertical air mass movement, convective activity and orographic lift; air mass movement over upwind mountain barriers; interception; gauge exposure; and the partial depletion of available condensation through precipitation induced by upstream barriers. Including the latter term yielded the highest correlation coefficient between measured winter precipitation and the orographic effect (r = 0.90). They concluded that long term average precipitation was not well correlated to station elevation except for points on the same ridge.

In coastal British Columbia orography is likely the principal lifting mechanism and we expect snowfall to increase with elevation. We note that an effective orographic component to precipitation is favoured by the following conditions: i) strong winds at right angles to mountain ridges, ii) air masses moist in depth containing existing cloud, iii) a large value of $\underline{h}$, the elevation at which the air mass ceases to rise. These conditions are usually best met in areas experiencing cyclonic or frontal conditions, such as coastal British Columbia (Table 3.2). Our search for a simple predictive model has concentrated on elevation.

Woo (1972) noted little correlation between rainfall and elevation at the UBC Research Forest (Haney, B.C.) over elevations ranging from 0-50 m when precipitation was <10 mm. As precipitation increased at the Research Forest, the orographic component was observed to increase. These findings are similar to those of Walkotten and Patric (1967) near Hollis, Alaska. Hetherington (1976) examined the orographic component of rainfall at the Research Forest in detail. He found significant correlations between amount of orographic rainfall and wind speed normal to the mountain barrier, moisture content of lower atmosphere, freezing level, and air mass stability. The dominating factor was topographic configuration.

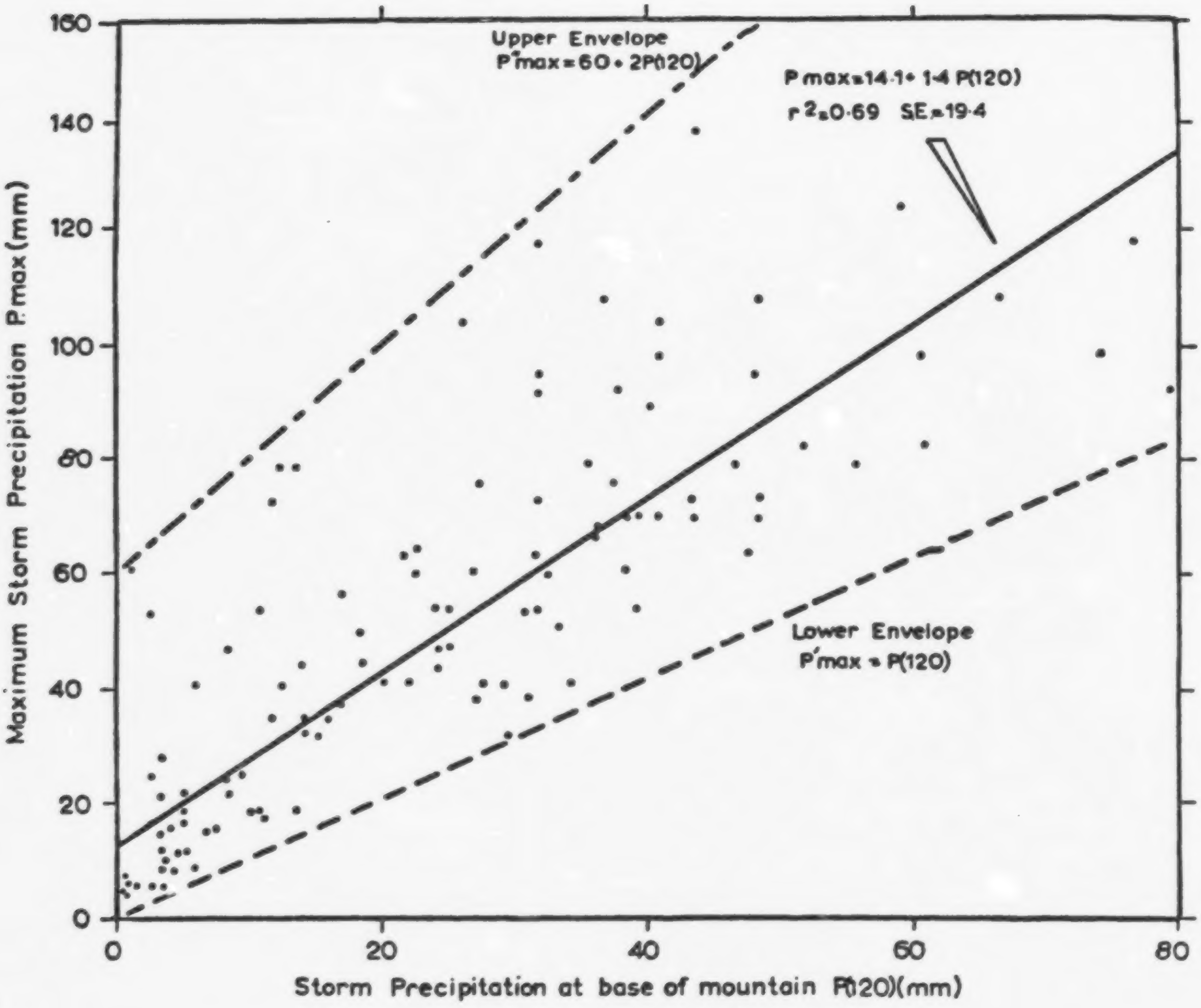


Figure 3.8 Maximum precipitation recorded for storms on Mt. Seymour as a function of storm precipitation at the base of the mountain during the winters of 1969-70 and 1970-71. The distance of the regression line from the lower envelope represents the orographic effect (from Fitzharris 1975: 234).

On Mt. Seymour, Fitzharris (1975) also noted stronger orographic effects when precipitation was greater (Fig. 3.8). The simple regression representing his relationship is:

$$\underline{P}(\underline{max}) = 14.1 + 1.4\,\underline{P}(\underline{120}) \qquad (3.1)$$

$$(r^2 = 0.69)$$

where $\underline{P}(\underline{max})$ = maximum storm precipitation (mm), and $\underline{P}(\underline{120})$ = precipitation (mm) at 120 m (the base of Mt. Seymour). Following Elliott and Shaffer (1962) the amount by which storm precipitation exceeds that at the base of the mountain (lower envelope of Fig. 3.8) is termed the orographic component of precipitation: $\underline{P}(\underline{oro}) = \underline{P}(\underline{max}) - \underline{P}(\underline{120})$. In other words, as storm size increased on Mt. Seymour, the orographic component also increased (Eq. 3.1, Fig. 3.8).

Table 3.3 presents Fitzharris' (1975) multiple stepwise regression equations relating precipitation at any elevation $\underline{H}$ to precipitation at the base of the mountain. All regressions were statistically significant ($P \leq 0.01$). The orographic component was greatest for occluded fronts and least for Arctic air associated with maritime fronts (Table 3.3). These empirical observations correspond well with the theoretically based conclusions of Sawyer (1956) and Williams and Peck (1962).

The effect of the orographic component is that total precipitation increased with elevation on Mt. Seymour during both the winters of 1969-1970 and 1970-1971 (Fig. 3.9). The

Table 3.3 Regression equations estimating storm precipitation (P) as a function of elevation (H) on Mt. Seymour (data of Fitzharris 1975)[a].

Data base	n	Equation	r^2	SE
Winter 1969-70	627	$P(H)$[b] $= -2.8 + 1.05\,P(120) + 0.010H$	0.81[c]	9.7
Winter 1970-71	748	$P(H) = -9.7 + 1.21\,P(120) + 0.024H$	0.78	13.2
Both winters	1375	$P(H) = -6.8 + 1.15\,P(120) + 0.017H$	0.79	12.1
Elevations below 870 m	1375	$P(H) = 2.91 + 0.86\,P(120) + 0.0005\,P(120)\,H$	0.89	7.8
Elevations above 870 m	1375	$P(H) = 4.82 \pm 0.47\,P(870) + 0.0005\,P(870)\,H$	0.78	14.3
Maritime cold fronts	168	$P(H) = -7.4 + 1.24\,P(120) + 0.016H$	0.82	11.7
Maritime fronts with warm sector	160	$P(H) = -7.2 + 1.21\,P(120) + 0.017H$	0.77	14.3
Maritime occluded fronts	424	$P(H) = -8.6 + 1.17\,P(120) + 0.021H$	0.80	11.5
Non-frontal systems	200	$P(H) = -4.6 + 1.04\,P(120) + 0.018H$	0.68	12.9
Arctic fronts associated with maritime fronts	48	$P(H) = -0.9 + 1.09\,P(120) + 0.007H$	0.81	11.0

[a] Data from 57 storms in winter 1969-70 and 68 storms in winter 1970-71. Includes a few cases where two consecutive storms were sampled as one.

[b] $P(H)$ = storm precipitation (mm) at some elevation H, where $120 < H < 1260$ m
$P(120)$ = storm precipitation (mm) at base of mountain (120m)

[c] All regressions have slope significantly > 0.00, $\underline{P} < 0.01$

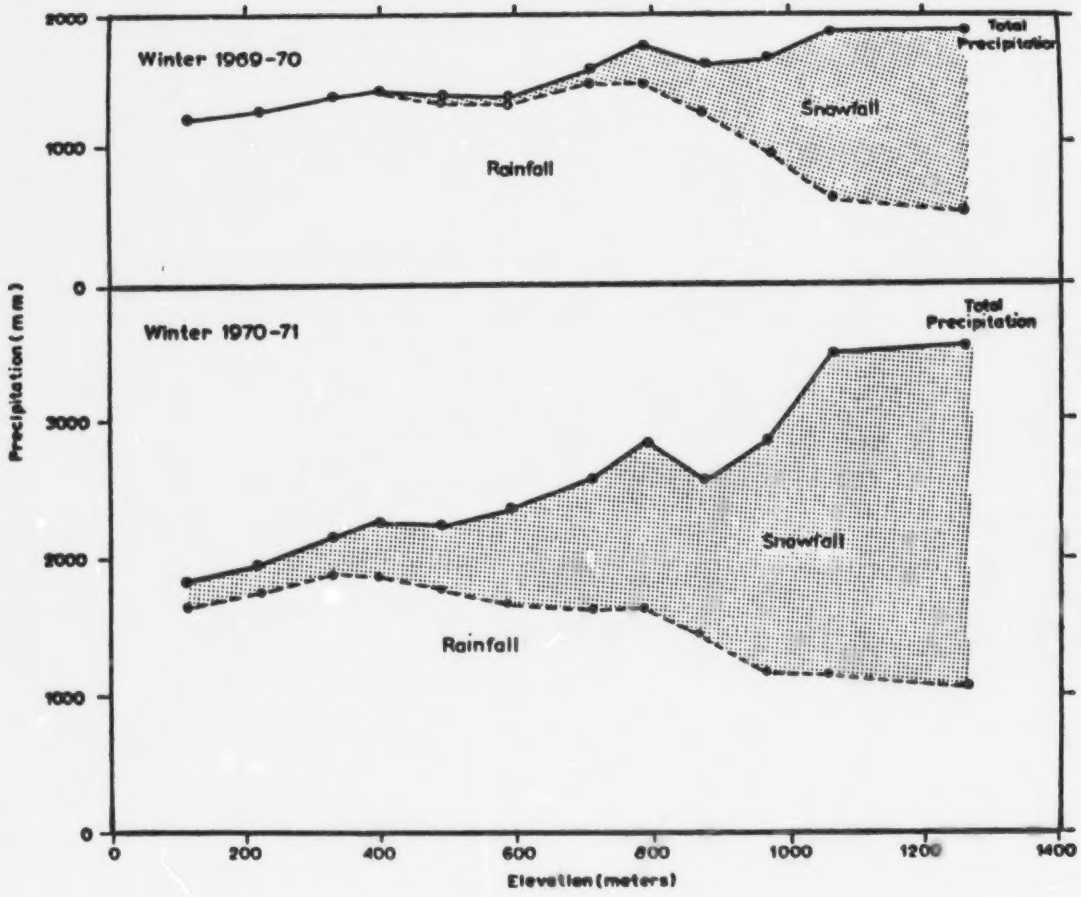


Figure 3.9 Variation in winter precipitation with elevation on Mt. Seymour during 1969-70 and 1970-71. When mean storm sizes were larger and precipitation was greater (1970-71) the orographic effect was larger (from Fitzharris 1975: 134).

slope of P(max) versus elevation was much greater in 1970-71 suggesting important year-elevation interactions. Reasons for the interaction appear straightforward and relate to the broad movements of air masses. During 1969-70 the Aleutian low was greatly expanded and more intense, producing persistent southerly flows over Mt. Seymour in what is normally the snowiest part of the year. During 1970-71 the Aleutian low was unusually weakened or eliminated by an extensive ridge in the eastern Pacific. A persistent northerly flow developed along with frequent outbreaks of Arctic air to the Pacific coast. These broad patterns produced a winter milder and drier than normal in 1969-70 and wetter than normal in 1970-71 (data from Vancouver International Airport). The mean storm size was much greater in 1970-71 (partially associated with Arctic air masses) thereby enhancing the orographic component. The available data thus suggest that any generalization concerning the effect of elevation on total seasonal precipitation must be tenuous. When storm sizes were greater, and total precipitation was greater, the orographic effect also was greater. The lack of greater generality is a direct result of the variable nature of air masses over the coast (Fig. 3.4).

Barry and Chorley (1974) suggested that the elevation of maximum precipitation should be at the level of the cloud base, because particles falling through drier air below would evaporate and lose mass. Fitzharris (1975) noted that the cloud base on Mt. Seymour was about 500 m, too low for that process to be operating frequently. He reviewed the work of several researchers in south coastal British Columbia who had suggested higher winter precipitation at mid-elevations (Walker 1961, Orloci 1964, and Wright 1966). Fitzharris pointed out probable empirical errors in their results (principally the use of snowpack measurements of assumed density) and concluded that precipitation increased monotonically with elevation at least to 1260 m. Other studies in south coastal British Columbia, including the Beaufort Range of Vancouver Island, have found a monotonic increase in both snowfall and total pecipitation (snow and rain) with increasing elevation on any particular slope (Schaefer and Nikleva 1973, Ferguson et al. 1974a and b).

3.2.2 Proportion of Precipitation Falling as Snow

One standard meteorological approach to prediction of snowfall vs rainfall involves determination of the thickness of the 500-1000 mb layer as determined by radiosonde ascents. The rationale is that thickness values for this layer vary with temperature and are therefore thought to reflect snowfall and rainfall processes. The approach has been used by Lamb (1955), Wagner (1957), Younkin (1967), and Tyner (1972). None of these studies referred to the northwest coast. Ranahan and Alexander (1979) tested the efficiency of the 500-1000 mb thickness for Vancouver Island and found it to be an unreliable predictor of snowfall versus rainfall. Much better was the temperature at 850 mb and a two-factor function utilizing temperature at 850 mb and surface wind speed (T850 and V of Fig. 3.10). The latter approach yielded correct forecasts in 77% of the cases. The approach does not, however, provide flexibility for prediction at various elevations and requires rare radiosonde data. It does indicate the importance of wind velocity noted earlier.

Hydrologically oriented researchers (e.g., Willen et al. 1971, Leaf and Brink 1973, and Fitzharris 1975) seem to have felt that prediction of snowline, the elevation at which precipitation falls as snow, could be explained reasonably well by temperature alone. Fitzharris (1975) tried to determine freezing levels on Mt. Seymour from data of radiosonde ascents at Port Hardy but found the predictions unreliable. They were incapable of predicting the orographic effect. Much better results were obtained by utilizing temperatures at the base of the mountain and a mean lapse rate of 7°C·km^{-1}.

Fitzharris defined "equivalent elevation" $\underline{H}(\underline{e})$ as the elevation at which all precipitation fell as snow and $\underline{H}(\underline{o})$ as the elevation of the new but incomplete snowline. Both $\underline{H}(\underline{e})$ and $\underline{H}(\underline{o})$ were closely related to $\underline{H}(\underline{fl})$, the mean height of the freezing level during the storm. Regression analysis yielded the following relationship (all elevations in m):

$$\underline{H}(\underline{e}) = 124.8 + 0.98\ \underline{H}(\underline{fl}) \qquad (3.2)$$

$$(r^2 = 0.82)$$

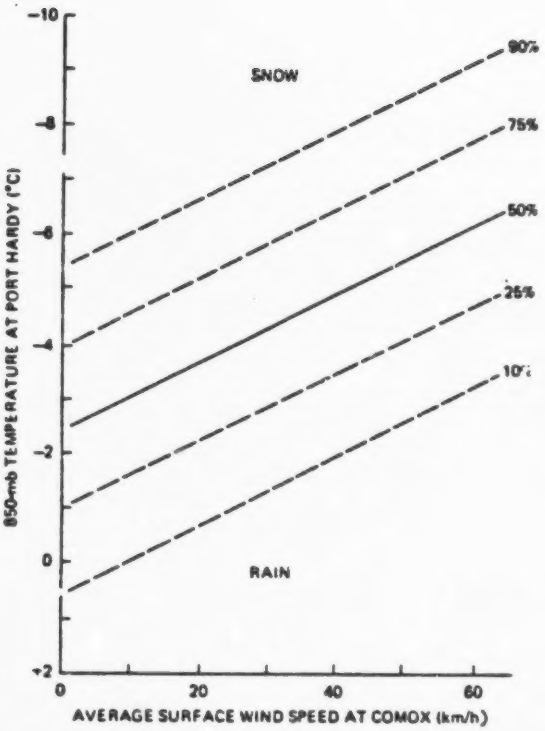


Figure 3.10 Conditional probability of snow at Comox as a function of 850 mb temperature at Port Hardy and the average surface wind speed at Comox. These probabilities are derived from the discriminant function X = T850 + 0.062(V) (from Ranahan and Alexander 1979: 12).

Freezing levels themselves varied with storm type depending on the air mass involved (Table 3.4). The 95% confidence interval was large (± 284 m, Fig. 3.11) due primarily to fluctuations in mean freezing level within storms (Table 3.4).

The position of $\underline{H}(\underline{o})$ was related to freezing level as:

$$\underline{H}(\underline{o}) = -79.8 + 0.92\,\underline{H}(\underline{fl}) \qquad (3.3)$$

$$(r^2 = 0.81)$$

The percent precipitation falling as snow in the zone in between $\underline{H}(\underline{e})$ and $\underline{H}(\underline{o})$ was determined by simple interpolation. When the freezing level fluctuated wildly, effective predictions were obtainable only by subdividing the storm.

The regression equations 3.2 and 3.3 are empirical and apply solely to the mesoscale area of Mt. Seymour. Note that predictions of both snowlines had slopes of very nearly 1.0. That implies that the lapse rate was constant. Validity of the equations over larger areas of coastal British Columbia is unknown. Topographic features and location doubtlessly will influence the parameters. The equations do represent our best available estimates for the predictions of snowlines within a storm.

3.3 Formation of Ice Crystals

The third and finest scale of processes involved in the creation of snow involves the microphysics of clouds. These

Table 3.4 Variation of mean freezing level of snow storms among synoptic storm types. Data of Fitzharris (1975) for winters 1969-70, 1970-71 combined.

Storm type	Mean ± S.D. (m amsl)	Nature of frequency distribution
Storms associated with Arctic air	295 ± 225	Positively skewed distribution with the mode at 100 m
Non-frontal storms	589 ± 292	Bimodal with peaks at 200 m and 600 m
Maritime cold front	729 ± 361	Bimodal with peaks at 200 m and 900 m
Maritime occluded fronts	770 ± 264	Normally distributed
Maritime fronts with a warm sector	891 ± 179	Normally distributed

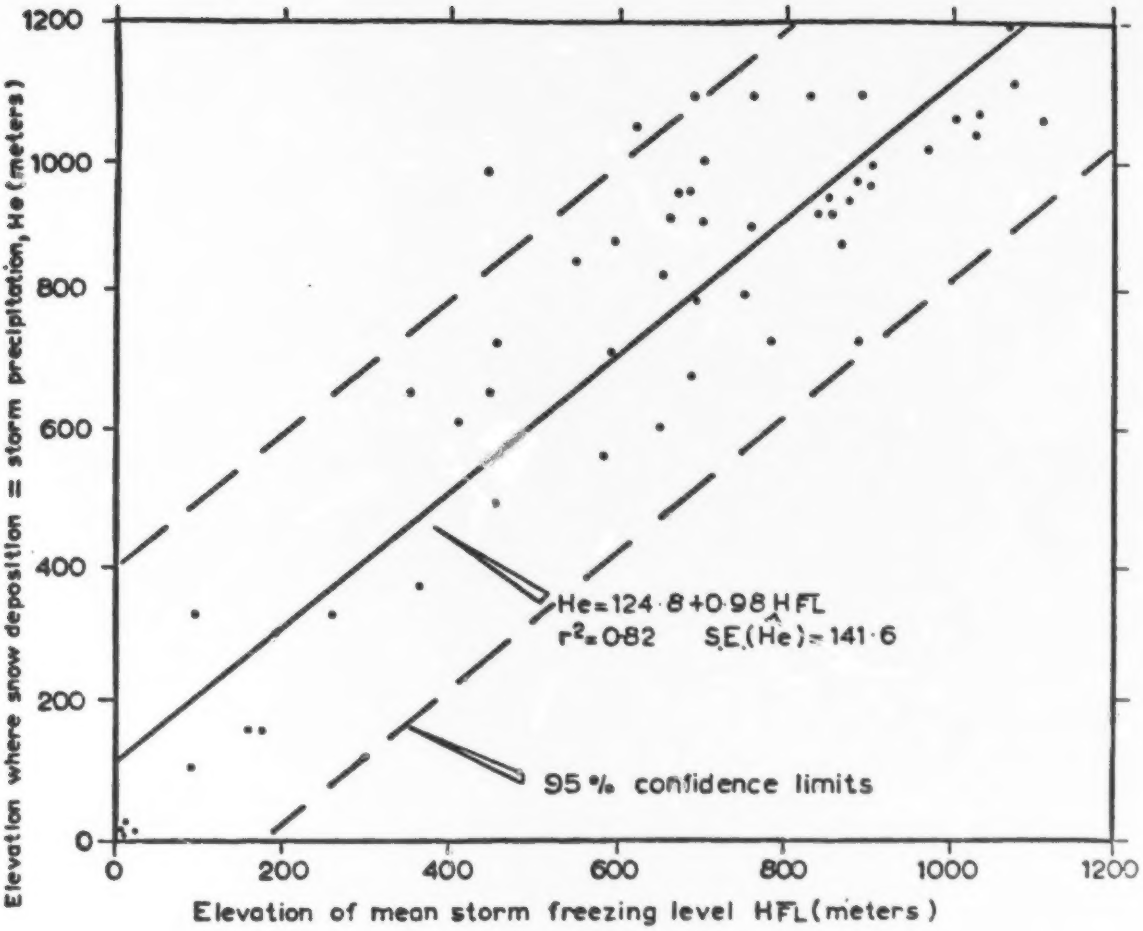


Figure 3.11 Relationship between the equivalent elevation and mean storm freezing level on Mt. Seymour during winters of 1969-70 and 1970-71 (from Fitzharris 1975: 255).

processes determine whether ice particles will form and what kind of particles will be formed. In coastal British Columbia, whether ice particles form depends on the speed of the processes acting (relative to the duration of the orographic effect just noted, Ch. 3.2), and on the presence of sufficient ice-forming nuclei. Moisture is unlikely to be limiting (Fig. 3.5), nor is temperature at higher elevations during winter (Fig. 3.1). What kind of particles form depends on the motion of air within clouds and relationships governing growth and aggregation of ice particles. These relationships are not yet firmly established (reviews of Mason 1971 and Pruppacher and Klett 1978). Moreover, the large variations in the concentrations and properties of atmospheric aerosols, and the great complexity of atmospheric motion make it difficult to construct a general theory of precipitation development, let alone apply such theory in a mountainous area. Our purpose here is to extract the few generalities we can.

Major empirical relations are illustrated (Fig. 3.12). The processes leading to the formation of ice crystals begin with some nuclei present in water vapour. That is because supersaturations as high as several hundred percent would be necessary for drop formation in homogeneous water vapour and such conditions do not occur in the atmosphere. Levels of supersaturation in the atmosphere remain below 10% and most often are below 1%. Drop formation in the atmosphere must occur via some process of heterogeneous nucleation involving aerosol particles. Those aerosol particles that are capable

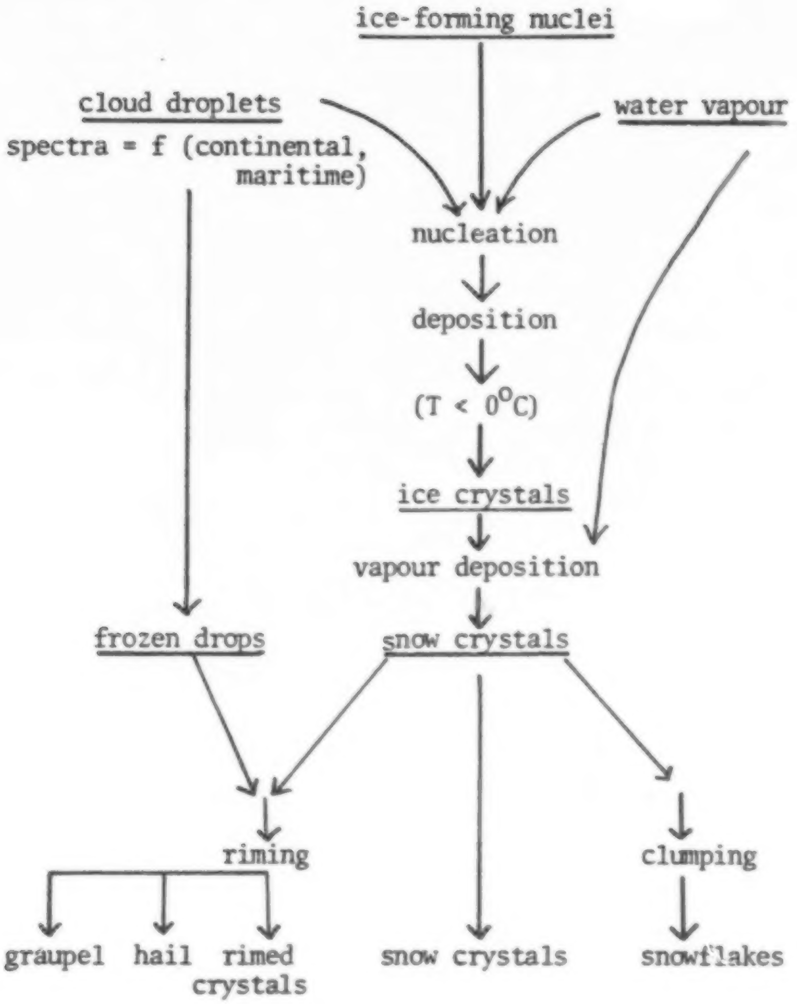

ice-forming nuclei
cloud droplets
spectra = f (continental, maritime)
water vapour
nucleation
deposition
(T < 0°C)
ice crystals
vapour deposition
frozen drops
snow crystals
riming
clumping
graupel
hail
rimed crystals
snow crystals
snowflakes

Figure 3.12 Simplified flow diagram of the formation of different types of 'snow'.

of initiating drop formation at the low supersaturations observed are termed cloud condensation nuclei (CCN). In the atmosphere the formation of ice crystals generally begins at temperatures too warm for the homogeneous freezing of water. Such behaviour indicates that some small fraction of the local aerosol particles also can serve as ice-forming nuclei (IN). That fraction is about one in 10^9 at -10°C.

3.3.1 Initiation of Ice Crystals

Cloud condensation nuclei (CCN).--Twomey and Wojciechowski (1969) provided a comprehensive review of CCN concentrations over various parts of the world. The broad findings of their review were:

1) There is no systematic latitudinal variation in concentration.

2) Continental air masses are generally richer in CCN than are maritime air masses.

3) The number concentration of CCN, N(ccn), increases as a power function of supersaturation (Sw: saturation ratio of moist air with respect to a plane water surface).

$$N(ccn) = aSw \qquad (3.4)$$

Considering the definition of CCN it is reasonable to assume that their concentration in a given air volume is to a large extent indicative of the drop concentration in a cloud which forms in that air volume. At the supersaturations which commonly occur in clouds (a few percent), number concentrations of CCN over continents typically range from 100 to 1000 cm^{-3}, while over oceans they range from a few tens to a few hundreds cm^{-3} (Twomey and Wojciechowski 1969). These values agree well with drop concentrations found in continental and maritime clouds (Squires and Twomey 1960, 1966, Braham 1968, Woodcock et al. 1971, Hindman et al. 1977). Abundance of CCN does not appear limiting in maritime clouds, these clouds simply have a broader drop spectra with larger drops (Fig. 3.12 and 3.19a). Indeed, a greater fraction of the total aerosol particles in maritime clouds act as CCN than in continental clouds. In air over land typically 1/100 or less (sometimes as large as 1/10) of the aerosol particles are capable of serving as CCN at supersaturations of 1% or less. Conversely, in air over oceans the fraction of CCN to total aerosol particles is considerably larger, ranging from 1/5 to 1/20 (review of Pruppacher and Klett 1978). Furthermore, the rate of activation of CCN relative to supersaturation may be greater in Maritime air. Values of k for equation 3.4 are frequently greater in maritime than in continental air masses.

Those aerosol particles that consist of water soluble, hygroscopic substances are most suitable for initiating the formation of water drops from water vapour, and therefore are most likely to act as CCN. It would seem reasonable that the greater rates of k (Eq. 3.4) and larger proportion of aerosol particles acting as CCN in maritime air are a function of the greater concentrations of sea salt particles (resulting from wave action bursting bubbles). Generally the concentrations of chloride particles (Cl^-) at ground level increase with increasing proximity to the ocean (Fig. 3.13).

Evidence available also suggests that even close to the ocean's surface sea salt particles account for less than one tenth of the total CCN. That conclusion is based on direct measurements of sea salt particle concentration and on indirect estimates of their rate of production and their residence time as reviewed by Pruppacher and Klett (1978). Generally, CCN concentrations in maritime and modified maritime air masses which have been overland less than 2 days rarely exceed 100 cm^{-3}, while concentrations in excess of 10^3 cm^{-3} are found in air which has been over land for several days (Figs. 3.13 and 3.14).

Conditions of coastal air masses are seldom stable. The concentration of all cloud condensation nuclei (CCN) may vary over several orders of magnitude within 24 h or less. Data from Figure 3.15 are from the Olympic Mountains in Washington and indicate that changes in air mass, wind speed, and direction all influence the CCN concentration. Twomey and Davidson (1970, 1971) noted that for a given location (Robertson, Australia) repeated patterns could be detected in diurnal variation of CCN concentration. Such may also be true

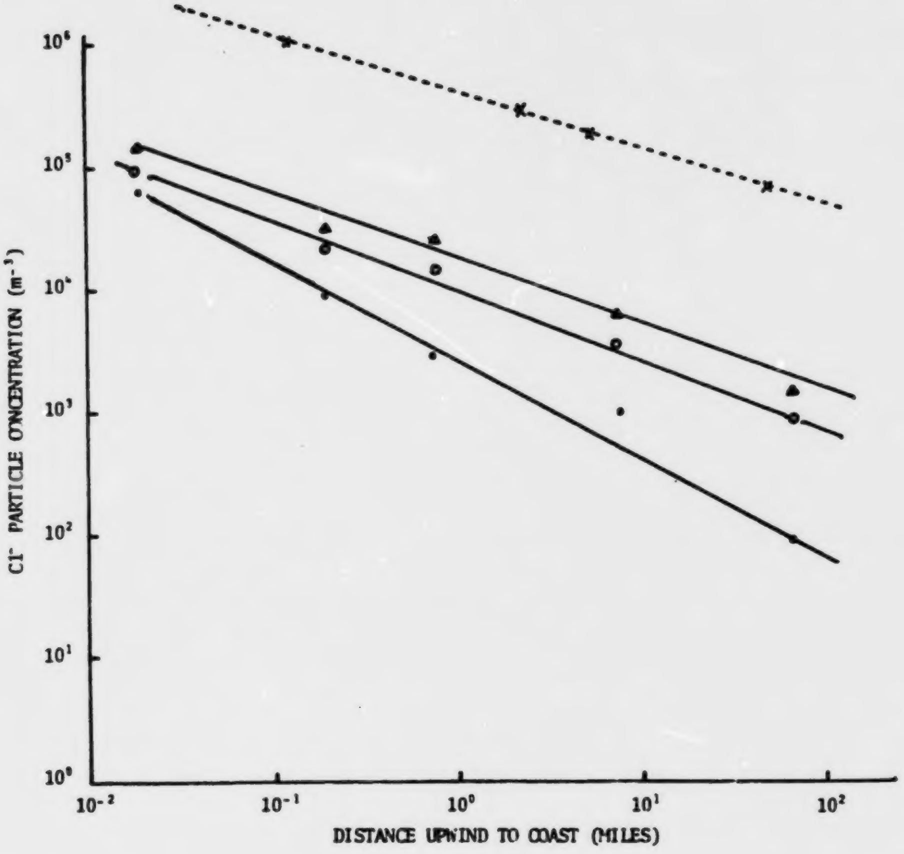


Figure 3.13 Variation in the concentration of chloride particles of various sizes at ground level as a function of increasing distance from sea shore. (x = Rossknecht et al. (1973) in Oregon ($r \geq 1.24$ μm); all others from Lodge (1955) in Puerto Rico, ▲ = $r \geq 3$ μm, o = $r \geq 6$ μm, • = $r \geq 10$ μm).

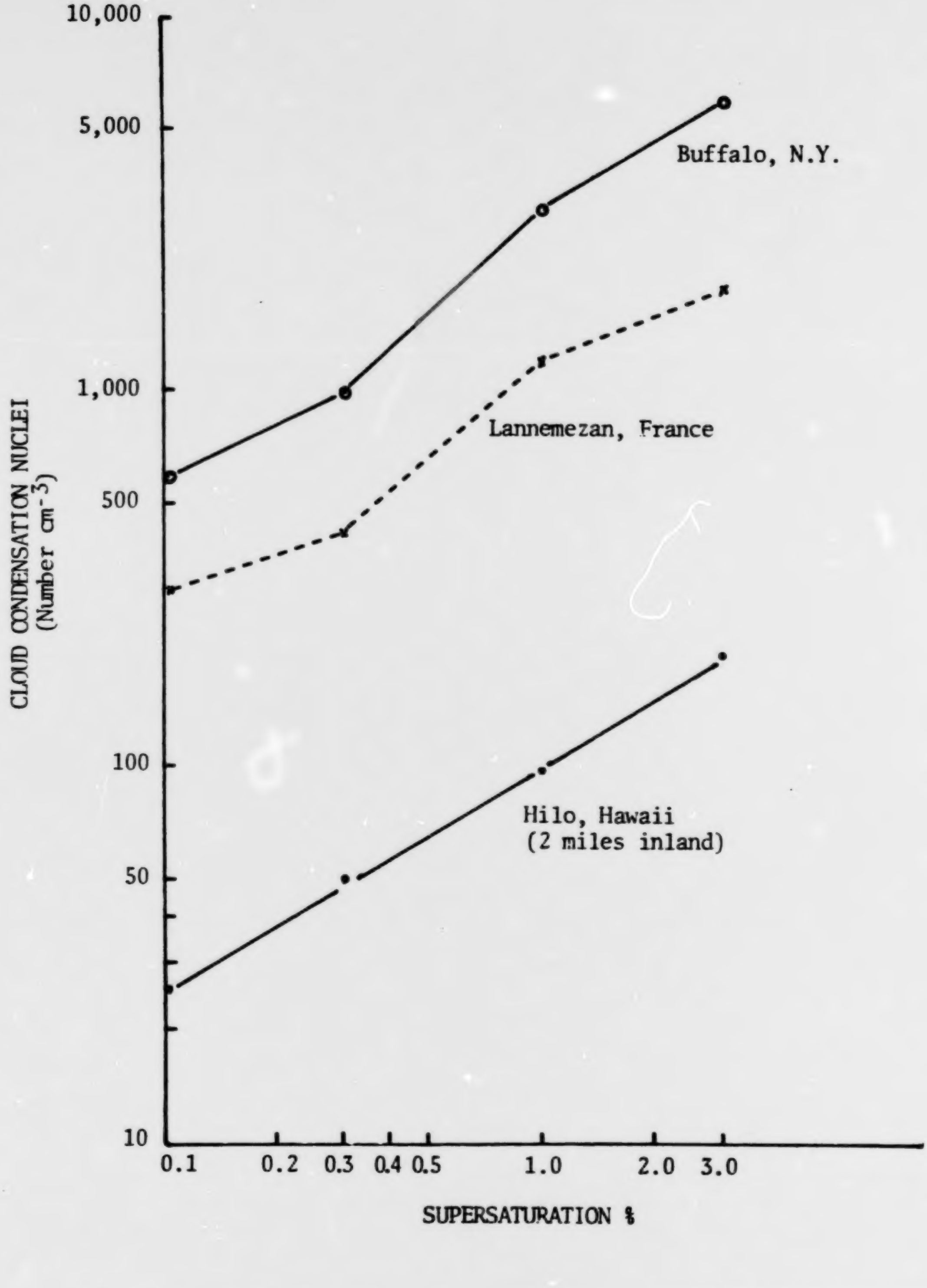


Figure 3.14 Variation in the concentration of cloud condensation nuclei required for activation at three locations as a function of supersaturation (data of Jiusto and Kocmond 1968: 103).

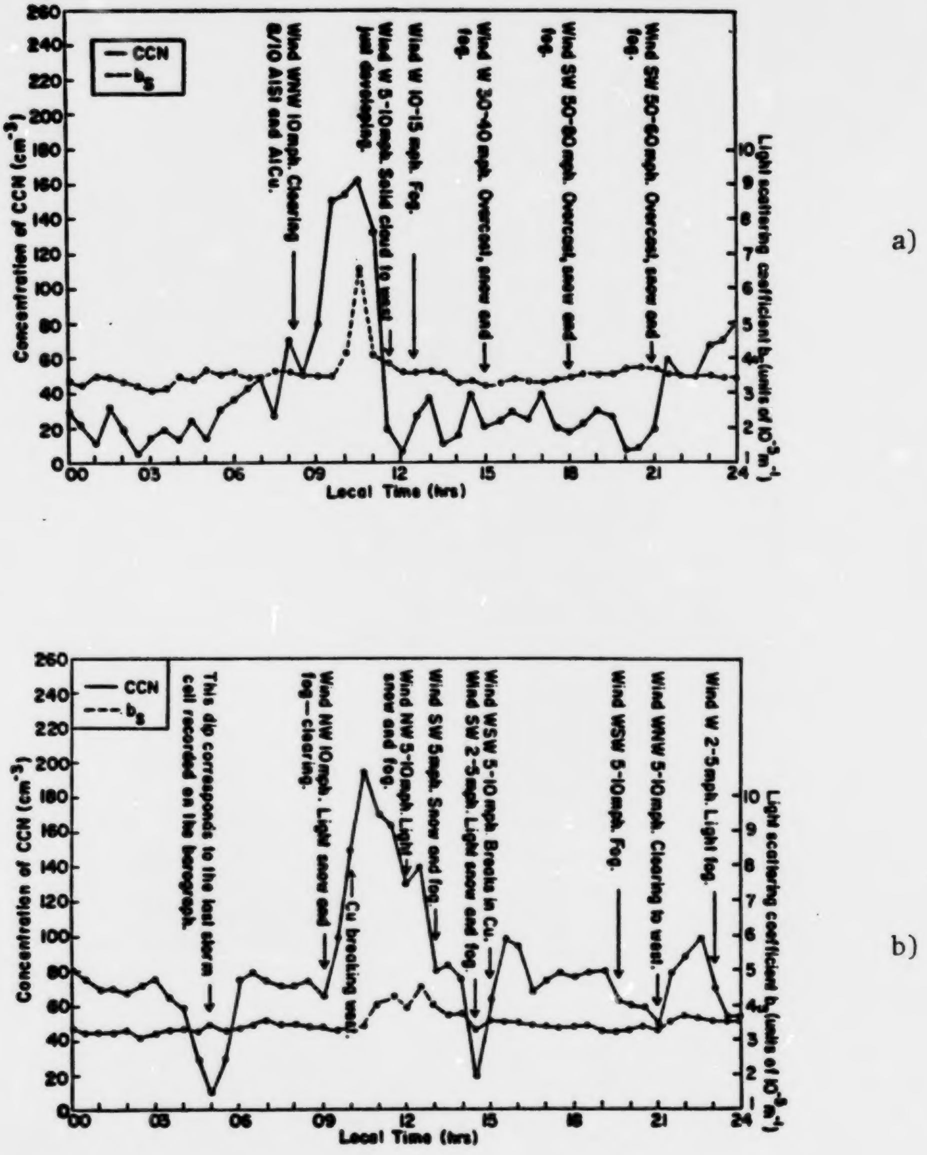


Figure 3.15 Variation with time in the concentration of cloud condensation nuclei activated at 1% supersaturation in air at 2025 m in the Olympic Mountains, Washington. The light scattering coefficient, b_S, is also illustrated.
a) 28 March, 1968,
b) 29 March, 1968 (from Radke and Hobbs 1969: 283).

of coastal areas subject to predictable on- and off-shore winds.

The variability in CCN concentrations makes it difficult to extract generalities, or to evaluate whether they are liable to prove limiting to coastal precipitation. Despite the lower concentrations in maritime air (Fig. 3.14), limitation appears unlikely or of short duration (Fig. 3.15). The variable but often higher rates of activation of CCN in maritime air apparently reduce the potential for limitation. The presence of hygroscopic salt particles may account, in part, for the higher rate of activition.

Ice-forming nuclei (*IN*).--Some CCN are capable of acting as ice-forming nuclei. Ice particles may form when cloud temperatures decrease below 0°C, but the probability that clouds will contain ice particles does not exceed the probability of containing only supercooled water until temperatures decline to about -12°C (Fig. 3.16). With further decreases in temperature the likelihood of ice particles being present increases such that at -20°C fewer than 10% of clouds consist entirely of supercooled water drops. Because the probability of the ice phase in clouds increases with decreasing temperature (Fig. 3.16), we might expect a monotonic increase in the concentration of ice particles with decreasing temperature. Such behaviour holds only in a minority of cases. More often, a rapid phase change to ice (*glaciation*) occurs within the range -4 to -25°C such that ice

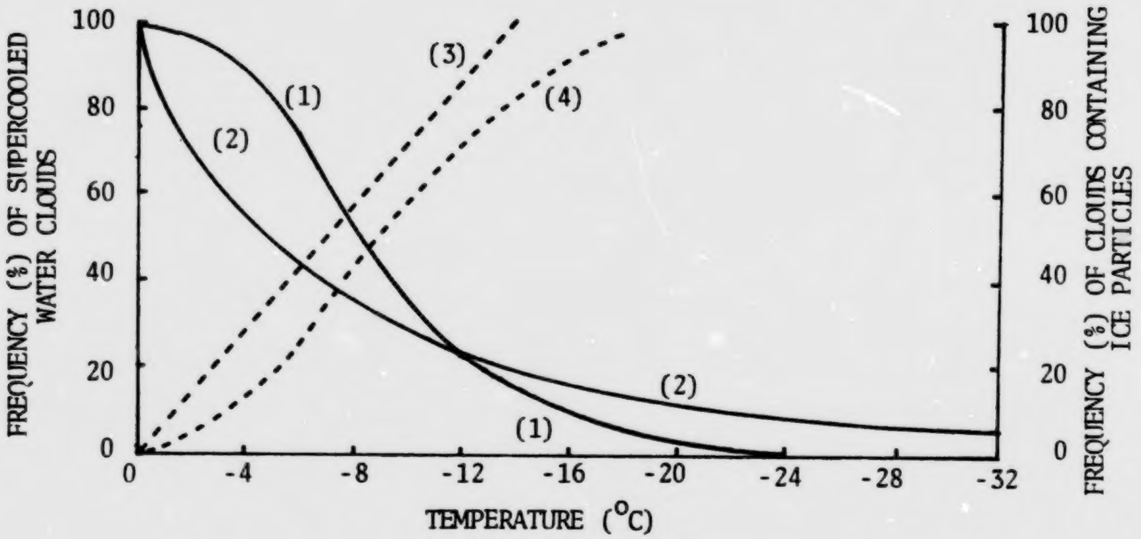


Figure 3.16 Frequency of occurrence of clouds containing supercooled water (———) and clouds containing ice crystals (-----) as a function of cloud top temperature. Data from: (1) Peppler (1940, in Pruppacher and Klett 1978), all-water clouds over Germany; (2) Borovikov et al. (1963), all-water clouds over European USSR; (3) Mossop et al. (1970), mixed clouds over Tasmania; (4) Morris and Braham (1968), mixed clouds over Minnesota.

particle concentration changes little with further decrease in temperature.

The apparently erratic response of glaciation appears to result from the manner in which aerosol particles initiate condensation. The production of ice-forming nuclei may proceed at relative humidities of less than 100% (with respect to water) as long as there is supersaturation with respect to ice. The number concentration of ice-forming nuclei increases with relative humidity at any given temperature (Fig. 3.17a) and correlates logarithmically with the supersaturation of ice independently of temperature (Fig. 3.17b). The relationship has the same form as equation 3.4 but incorporates supersaturation with respect to ice (Si) rather than water (Sw). The exponent or k, however, varies between regions (Huffman 1973). Furthermore, there are at least three modes by which ice-forming nuclei can initiate ice particle formation (Pruppacher and Klett 1978).

One result of this complexity is that number concentrations of snow crystals show no clear relationship with temperatures below about -5°C (Fig. 3.18). Fletcher's rule of Figure 3.18 states that number concentrations of ice-forming nuclei, IN, increase nearly exponentially with decreasing temperature (Fletcher 1962):

$$IN = a \exp (b \Delta T) \qquad (3.5)$$

where N is the number of ice-forming nuclei active at a

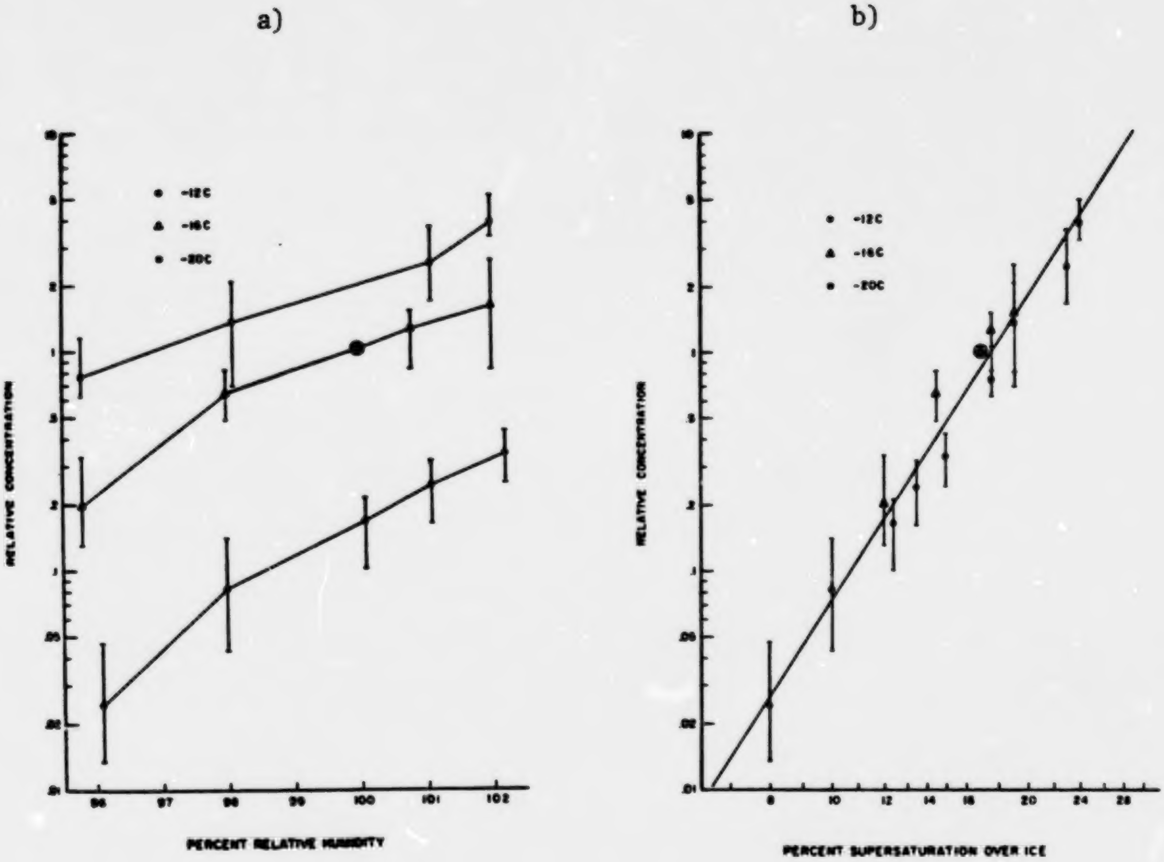


Figure 3.17 Concentration of ice-forming nuclei as a function of temperature and:
a) Relative humidity,
b) Supersaturation over ice (from Huffman 1973: 1081).

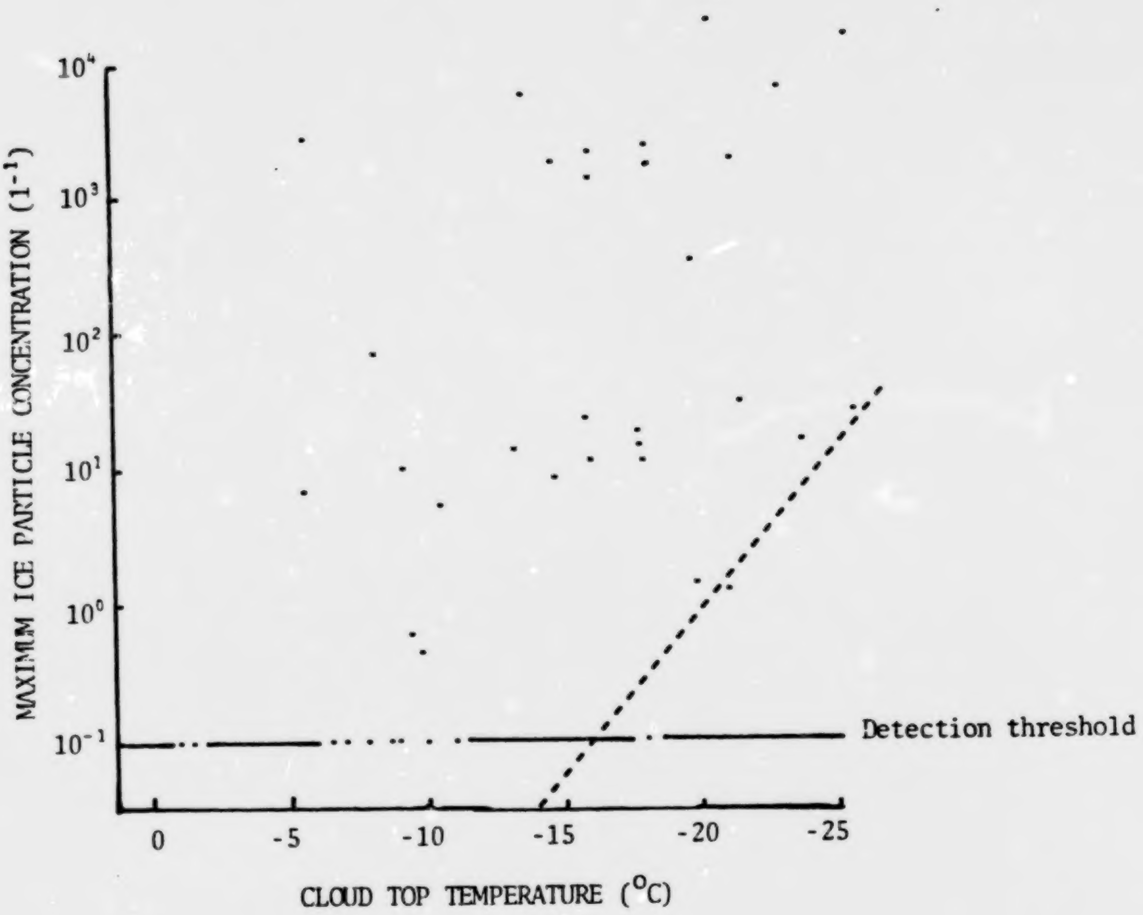


Figure 3.18 Maximum ice particle concentrations in clouds over the Cascade Mountains, Washington. Dashed line represents Fletcher's world wide, average ice nucleus concentration (modified from Hobbs et al. 1974b as presented by Pruppacher and Klett 1978: 41).

temperature warmer than T, $a = 10^{-5}\ l^{-1}$, $b = 0.6(^{\circ}C)^{-1}$, and $\Delta T = To - T$ (To is the melting temperature of ice). Data of Figure 3.18 suggest that Fletcher's rule provides only one boundary to the empirical pattern. Apparently factors other than temperature exert considerable influence on the formation of ice particles, at least over western coastal mountains. Relative humidity is a prime candidate, particularly given the moist air resulting from synoptic storm patterns (Section 3.1) and the fact that effects of supersaturation over ice are independent of temperature (Fig. 3.17b).

The major source of ice-forming nuclei is the earth's surface from which dust is raised by wind. Electron microscopy and electron diffraction techniques in a variety of areas have demonstrated that the central portion of snow crystals typically contain one solid silicate particle, usually identified as clay (kaolinite and montmorillonite) (Kumai 1951, 1957, 1961; Isono 1955; Isono et al. 1959, Kumai and Francis 1962, Rucklidge 1965). Hobbs et al. (1971b and c) used air mass trajectory analysis to show that high concentrations of ice forming nuclei over northwestern U.S. were often the result of local dust storms over the arid regions of northern China and Mongolia. Their empirical suggestions are born out by laboratory experiments which suggest that clay particles, particularly those from the northern hemisphere, become active as ice forming nuclei at relatively warm temperatures (-10 to -20°C, Schaefer 1950, Mason and Maybank 1958, Isono et al. 1959, Roberts and Hallett

1968). Moreover, Roberts and Hallett (1968) showed that clay particles, once having been involved in ice crystal formation, can exhibit considerably improved nucleability. For example, threshold temperatures for activity of kaolinite particles increased from -10 to -4°C after they had been activated.

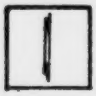
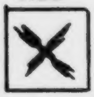
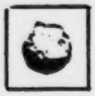
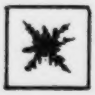
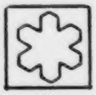
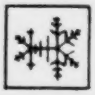
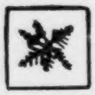
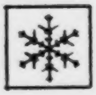
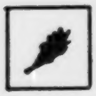
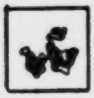
Superficially, these observations suggest that snowfall in coastal British Columbia depends on events in Mongolia. Two phenomena counter that suggestion. First, although fewer condensation nuclei are present in maritime air masses, those that are present appear to be activated more quickly than those in continental air. In their review Twomey and Wojcieschowski (1969) found k of Eq. 3.4 to be 0.5 in continental air, but 0.7 in maritime air. Second, there is an apparent but poorly understood "multiplicative effect" within stratiform clouds: the concentration of ice crystals exceeds the concentration of ice-forming nuclei by several orders of magnitude. The enhancement ratio is greater at cloud top temperatures of -10° to 0°C, declining to near zero at temperatures < -20°C (Hobbs 1974, Hobbs et al. 1974b, Heymsfield 1977). Hobbs et al. (1974b) collected data over the Cascade Mountains in Washington, Heymsfield's work included samples collected over Seattle.

Given the complexity of the processes involved, few unequivocal generalities can be applied to coastal British Columbia. Nevertheless, some broad implications are evident. Monotonic increases in concentrations of ice crystals and ice nuclei with decreasing temperature are unlikely to occur above coastal British Columbia (Fig. 3.18). They do occur in clouds over drier areas (e.g., Israel, Gagin 1971), and we interpret the erratic concentrations observed in the Cascade Mountains to imply that changes in relative humidity dominate the initiation of ice crystals over south coastal British Columbia. Available evidence suggests that clay particles, which can function actively at warm temperatures, may dominate among ice-forming nuclei. Given the moist nature of maritime air masses it is unlikely that either moisture or ice-forming nuclei are often limiting. Of the two, the abundance of ice-forming nuclei might be limiting more often. However, the enhancement ratios of ice crystals over ice-forming nuclei are greatest at warmer temperatures which prevail on the coast.

The major limitation in southwestern British Columbia is likely to be temperature or the rate of cooling of incoming warm air. The broad pattern predicted by empirical lapse rates was noted (Fig. 3.1). Rates of nucleation of ice crystals appear to be rapid, especially at higher levels of supersaturation (Pruppacher and Klett 1978). One potential limiting factor over southwestern British Columbia is thus sufficient time for the ice crystals to grow in air that is $\leq$ 0°C. Height of the coastal mountains, the empirical lapse rate ($7°C \cdot km^{-1}$), and moderate wind velocities ensure sufficient cooling under most winter conditions, but the air mass could pass so rapidly over the mountains that snow crystals could not grow and fall. Velocities frequently must be rapid enough to permit the presence of liquid water and riming.

Heymsfield (1977) found that vertical velocities in excess of 50 $cm \cdot s^{-1}$ at temperatures lower than -5°C were necessary for liquid water occurrence in deep, stratiform ice clouds. These conditions are unusual for warm frontal overrunning systems in which vertical velocity is typically low (2-10 $cm \cdot s^{-1}$) but could be common in the orographic lifting along coastal British Columbia. Velocities in warm frontal occlusions are higher than in overrunning systems, increasing from 1 to 20 $cm \cdot s^{-1}$ and may attain 55 $cm \cdot s^{-1}$ in convective regions (Heymsfield 1977). Hobbs et al. (1975) also found liquid water in the convective regions of a warm frontal occlusion at -5°C, but only light riming at the same temperature in regions of weak uplift. On Mt. Seymour 66% of all storms sampled (n = 200) produced snow (Table 3.1). Among the storm types occurring frequently, maritime occluded fronts exhibited the highest probability of producing snow (P = 0.69, Table 3.2). Riming occurred in at least 62% of the storms sampled (Table 3.5). Together, these data suggest that air velocity is usually sufficiently great to maintain liquid water content in the clouds (thus producing riming) and to ensure orographic uplift, but rarely so great that there is insufficient time for solid precipitation to form and fall (also see Ch. 3.3.2).

Table 3.5 Frequencies of snow crystal types in storm snowfalls on Mt. Seymour during the winters of 1969-70 and 1970-71 (data of Fitzharris 1975, App I).

Crystal type	C-Code sensu Magono and Lee (1966)	Percent occurrence in storms sampled
Needles - elementary and combinations	N1a  N1b N2a 	79
Lump graupel and graupel-like snow	R4b 	62
	 R3b	5
Crystals with broad branches	 P1c	45
Fern-like crystals	 P1f	29
Radiating assemblages of dendrites	 P7b	17
Ordinary dentritic crystal	 P1e	14
Rimed broken branches	 I3b	'sometimes'
Ice particles	 I	'sometimes'

3.3.2 Growth of Ice Crystals.

The preceding observations suggest that considerable ice particle formation in British Columbia occurs in relatively moist clouds at temperatures of -5 to -10°C. The conditions under which particle formation occurs determines the structure and mass of individual particles which in turn influence the character of the resulting snowpack.

There is a large literature dealing with the growth of ice crystals under experimental conditions, and with the theory of crystal growth acknowledging the chemical structure of water and underlying physics. Some theoretical and experimental work has been verified under natural conditions. Pruppacher and Klett (1978) provided the most comprehensive (714 pp) and recent review. Again, our purpose here is to summarize those general relationships that appear most relevant to conditions in coastal British Columbia.

Generally, when supercooled water drops coexist with ice particles, the latter grow at the expense of the former; either by water vapour diffusion to the ice particles (deposition) or/and by drops colliding with and freezing on ice particles (riming). Crystals which grow by riming must usually first grow by deposition. Ice particles which have grown only by deposition or vapour diffusion are termed ice or snow crystals. They have a single crystallographic orientation. Ice crystals may also grow by collision with other crystals, often referred to as clumping. Aggregates of snow crystals are termed snowflakes (Fig. 3.12). In 1951 the International Commission on Snow and Ice proposed a classification for solid precipitation. It is a simplified scheme representing the six main structural types of snow crystals, plus irregular crystals, graupel (heavily rimed crystals), ice pellets, and hail. More recently Magono and Lee (1966) extended the classification of natural snow crystals to 80 basic forms. LaChapelle (1980) estimated that 99% of snow crystals commonly observed in nature were accommodated by the latter system and it is the one used here.

The raw material upon which growth processes act differs between maritime and continental clouds. Drops within maritime clouds tend to be larger (Figures 3.12 and 3.19a). Braham (1968) used the frequency distributions of Figure 3.19a to derive the average diameters for his calculations of rates of drop growth (30 µm and 40 µm for continental and maritime clouds, respectively). Drops were predicted to grow faster in maritime clouds (Fig. 3.19b) which agrees with the observation that maritime clouds precipitate more often than do continental clouds (W of Fig. 3.19b represents liquid water content of the clouds).

Braham's calculations are more appropriate for rain and assume that drops grow only by collision and coalescence. However, the calculations do suggest that growth rates of precipitate will be relatively rapid in the cloud types common to coastal British Columbia. They help to reveal why orographic precipitation occurs on the windward side of

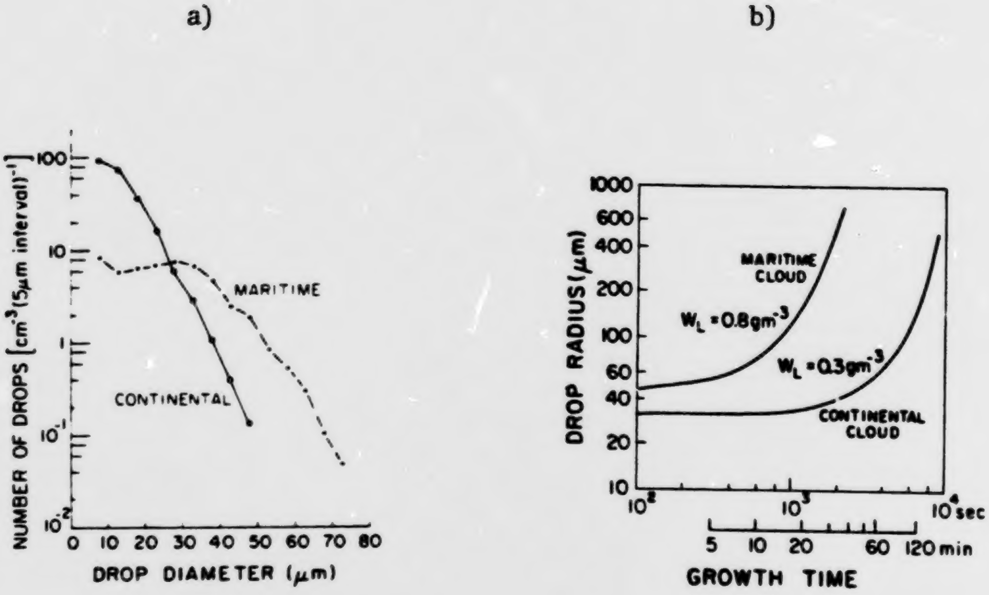


Figure 3.19 a) Frequency distributions of the size of cloud droplets in maritime-type (Caribbean Ocean) and continental-type (Southwestern U.S.A.) cumulus clouds.
b) Growth rates of drops growing by collision and coalescence in maritime and continental clouds having the droplet distributions illustrated in a) (both from Braham 1968: 344 and 345 as modified by Pruppacher and Klett 1978: 506 and 507).

coastal mountains even when wind velocity is great and h is attained quickly (Fig. 3.7).

Growth by vapour diffusion.--During the winter, the Wegener-Bergeron-Findeisen mechanism is more commonly operative than simple coalescence. Wegener (1911) showed through thermodynamic principles that at temperatures below 0°C supercooled water and ice-crystals cannot coexist in equilibrium. Bergeron (1933) proposed that precipitation was due to the colloidal instability in clouds containing both supercooled drops and ice crystals; Findeisen (1938) provided further supportive evidence. Although the physics are complicated, the process can be stated simply: ice crystals invariably grow by vapour diffusion at the expense of the supercooled drops until either all drops have been consumed or all ice crystals have fallen out of the cloud. The theoretical maximal growth rates invoking diffusion occur from -14 to -17°C and are about 3.0 to 5.5 x 10^{-8} $g \cdot s^{-1}$ (Fig. 3.20; Byers 1965). Most of the snowfall in coastal B.C. probably originates by this mechanism after which other growth processes become involved; specifically, accretion or riming and aggregation of individual crystals by collision (Fig. 3.12). Raindrops normally go through the same process at higher elevations, but subsequently melt before reaching the ground.

Laboratory experiments reveal that during the initial growth phase by diffusion the rate of growth along specific

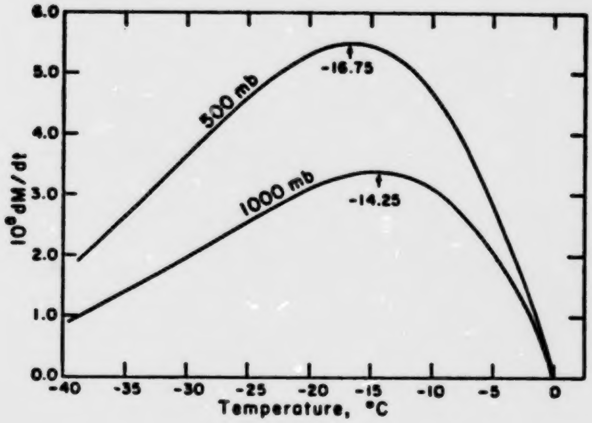


Figure 3.20 Influence of temperature on the normalized growth rate in mass (dm/dt) of an ice crystal growing by diffusion in a water-saturated environment (from Byers 1965: 123).

crystal axes varies with temperature and supersaturation in a characteristic manner. Laboratory findings of Hallett and Mason (1958), Kobayashi (1961), Weissweiler (1969), and Mason (1971) are summarized in Figure 3.21. They reveal that at large vapour density excess, the ice crystal shape changes with decreasing temperature from a plate to a needle, to a column (sheath), to a sector plate, to a dendrite, back to a sector plate, and finally back to a column. The cyclic change in shape results because the preferential growth along crystallographic axes shifts at temperatures near -4°, -9° and -22°C. The first two transition temperatures are sharply defined, the latter may occur over a range of several degrees but is centered around -22°C. In contrast, at very low vapour excess crystal shape changes between a short thick column and a thick plate near -9°C and -22°C (Fig. 3.21). Close to ice saturation, crystal shape does not vary with temperature but assumes the equilibrium shape of a thick hexagonal plate.

A large number of observations throughout the world document that natural snow crystals exhibit the same characteristic changes in shape with temperature and humidity. Magono and Lee (1966) summarized 18 studies of natural crystals in snow-forming clouds (Fig. 3.22) which agree closely with laboratory observations (Fig. 3.21). Pruppacher and Klett (1978) stated that seven more recent studies, including three in Washington by Hobbs et al. (1971a, 1972, 1974a) confirm the agreement. Given these underlying relationships with temperature and humidity, we would expect a

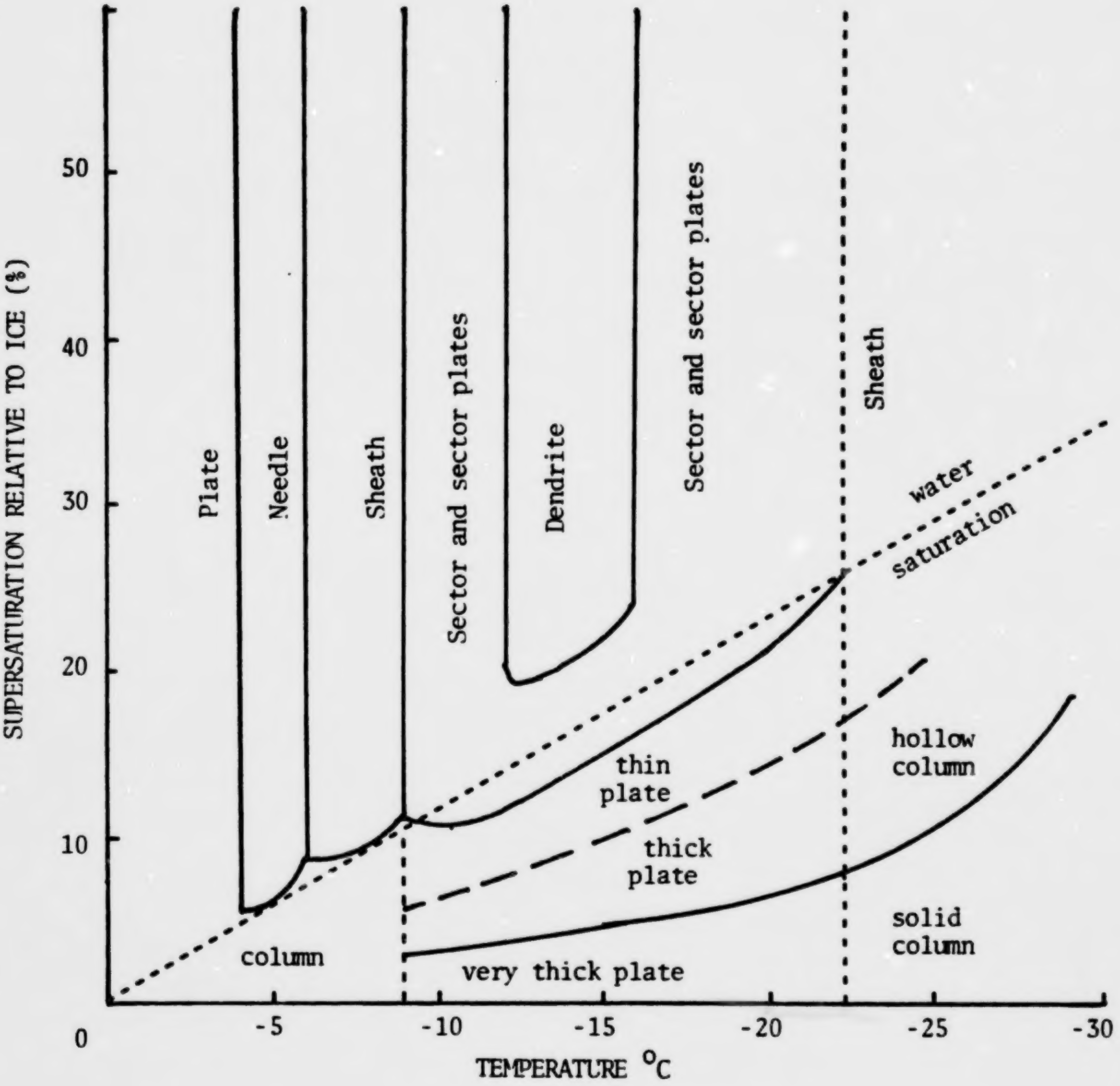


Figure 3.21 Variation of ice crystal form as a function of temperature and supersaturation (based on laboratory observations of Hallett and Mason 1958, Kobayashi 1961, Weissweiler 1969, and Mason 1971). Relationships illustrated are a revised presentation of the classic 'Nakaya diagram' (Nakaya 1954).

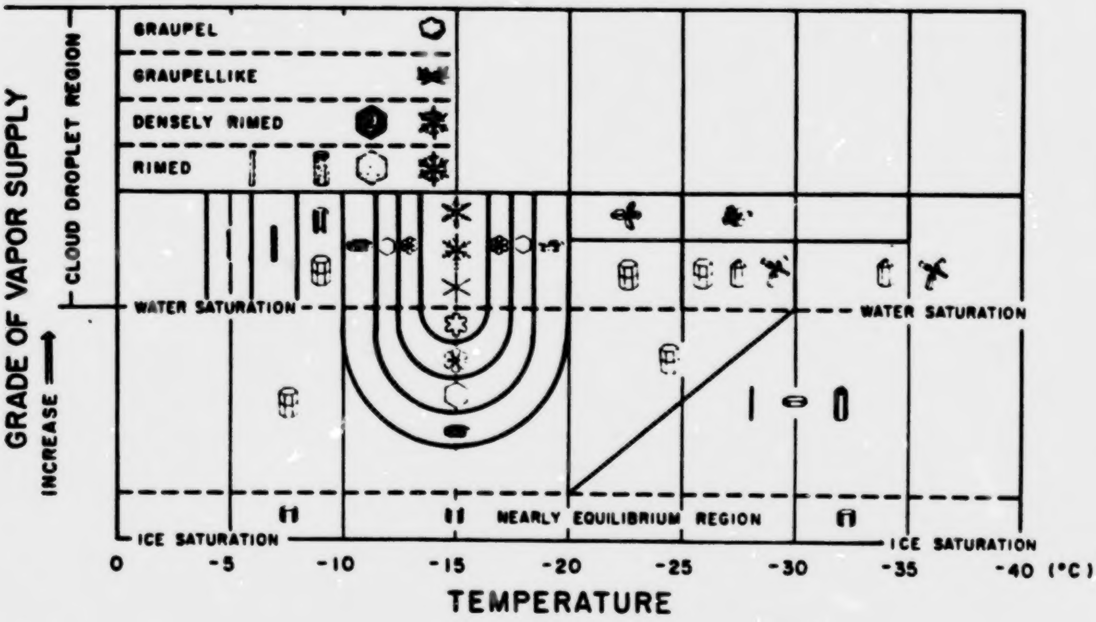

GRAUPEL
GRAUPELLIKE
DENSELY RIMED
RIMED
CLOUD DROPLET REGION
WATER SATURATION
WATER SATURATION
NEARLY EQUILIBRIUM REGION
ICE SATURATION
ICE SATURATION
GRADE OF VAPOR SUPPLY
INCREASE
0
-5
-10
-15
-20
-25
-30
-35
-40 (°C)
TEMPERATURE

Figure 3.22 Temperature and humidity conditions associated with the form of natural snow crystals of various types (from Magono and Lee 1966).

predominance of rimed crystals (with needles and graupel most evident) in the generally moist and warm conditions of coastal forests (Figs. 3.21, 3.22). Data of Fitzharris (1975) meet these expectations (Table 3.5). Fitzharris noted that in almost all cases crystals were rimed. Needles occurred most frequently as is expected for temperatures around -3 to -8°C; graupel was common, as is expected at moderately low temperatures (-15°C or warmer) with water vapour abundant.

There are few data documenting sizes of snow crystals in the Pacific Northwest; those available are primarily from the Cascade Mountains of Washington (e.g., Hobbs et al. 1974a, Locatelli and Hobbs 1974). Mean sizes are about 1.0 mm and 2.5 mm for maximum dimensions of needles and dendrites respectively (Fig. 3.23). Unfortunately, the data of Hobbs et al. (1972) are normalized (Fig. 3.23a) and we cannot compute relative frequencies of crystal types to compare with data collected by Fitzharris (1975) on Mt. Seymour. Further data on the dimensions and mass of solid precipitation particles in the Pacific Northwest are summarized in Table 4.1.

Data of Hobbs et al. (1974a) document the width:length relationships of needles in clouds over the Cascade Mountains. Needles (N1a) attained diameters of 20 to 70 μm and lengths of 200 to 1500 μm (Fig. 3.23b). Laboratory data of Middleton (1971) document rates of growth by diffusion for different crystal forms (Fig. 3.24). We see from Figure 3.24b that an N1a crystal growing by diffusion at water saturation and -18°C attains a length of 1500 μm (the largest value observed over

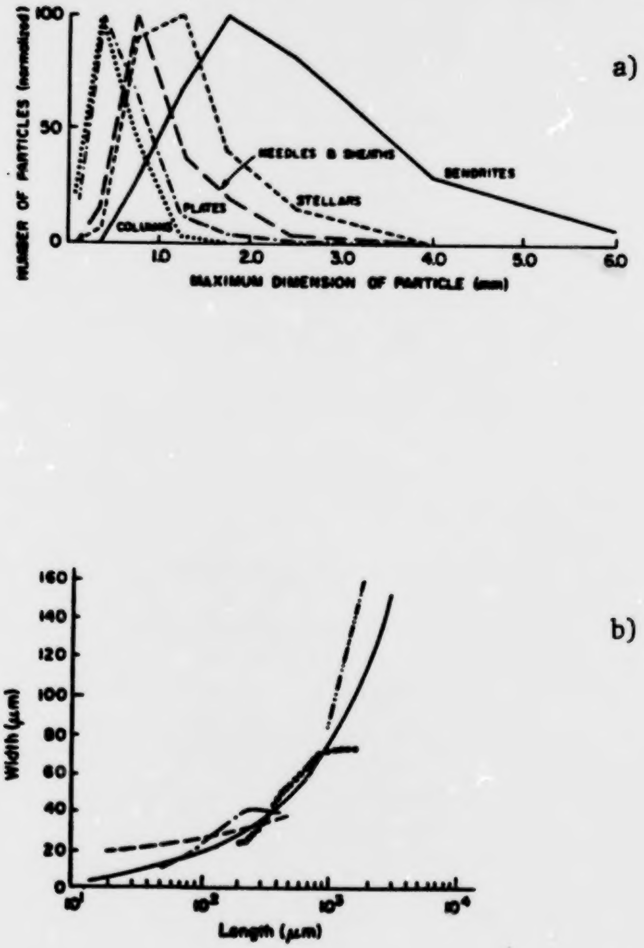


Figure 3.23 a) Size distributions of snow crystals collected at Alpental base and the Hyak and Kecheelus Dam, Cascade Mountains, Washington (from Hobbs et al. 1972 in Pruppacher and Klett 1978: 39).
b) Dimensions of needles observed in clouds. Data are from Magono (1954) —··— ; Isono (1959) ----- ; Ono (1969) —·— ; Auer and Veal (1970) ——— ; and Hobbs et al. 1974a ····· (from Hobbs et al. 1974a: 2200).

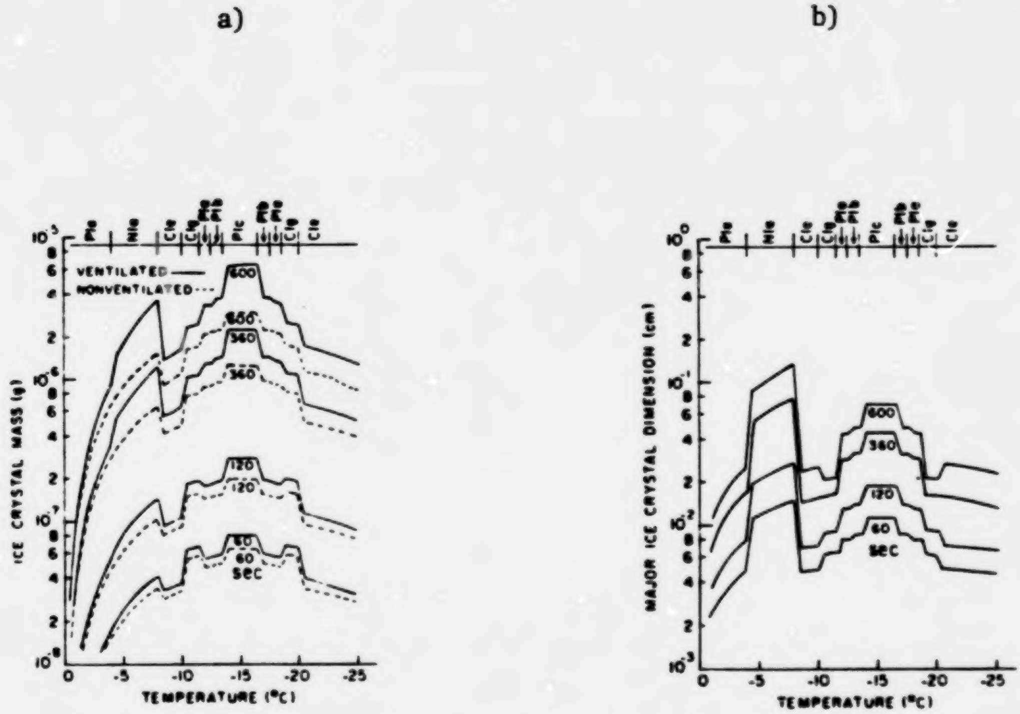


Figure 3.24 a) Mass of ice crystals grown by vapour diffusion in a water-saturated environment as a function of time (sec) and temperature, including (solid line) and excluding (dashed line) ventilation, at 700 mb.
b) Major dimension of ice crystals grown by vapour diffusion in a water-saturated environment as a function of growth time (sec) and temperature, at 700 mb (both from Middleton 1971, in Pruppacher and Klett 1978: 455).

the Cascade Mountains) within 600 sec or 10 min. Crystals with broad branches were also common on Mt. Seymour (Table 3.5). The mean major dimension of such crystals over the Cascade mountains was about 1 to 2 mm. From Figure 3.24b we see that a P1c crystal growing by diffusion at water saturation and -15°C attains a diameter of 4 mm in 6 minutes. Growth during ventilation (Fig. 3.24a) simply accounts for the effect of ventilation on diffusion of water vapour and heat once the crystal has grown to sufficient size by vapour diffusion that it has appreciable fall velocity.

Growth by aggregation.--Once ice or snow crystals have sufficient mass (Fig. 3.24a) they fall. Under certain cloud conditions snow crystals collide and aggregate to form snowflakes. More than one half the mass of the solid precipitation that reached the ground near the crest of the Cascade Mountains in the winter was in the form of aggregates (Hobbs et al. 1974a). It is thus important to review the relationships underlying aggregation. Whether growth of a crystal is caused by collision with frozen drops (riming) or aggregation with other snow and ice crystals, the rate of increase in mass of the ice crystal is dependent on its mass, dimensions, and fall speed. These are covariate with each other (Table 4.1). One empirical pattern dominates, the relationship with air temperature (Fig. 3.25).

Dobrowolski (1903) observed 283 aggregation episodes, of which 83% occurred between +1 and -5°C, 9% between -5 and

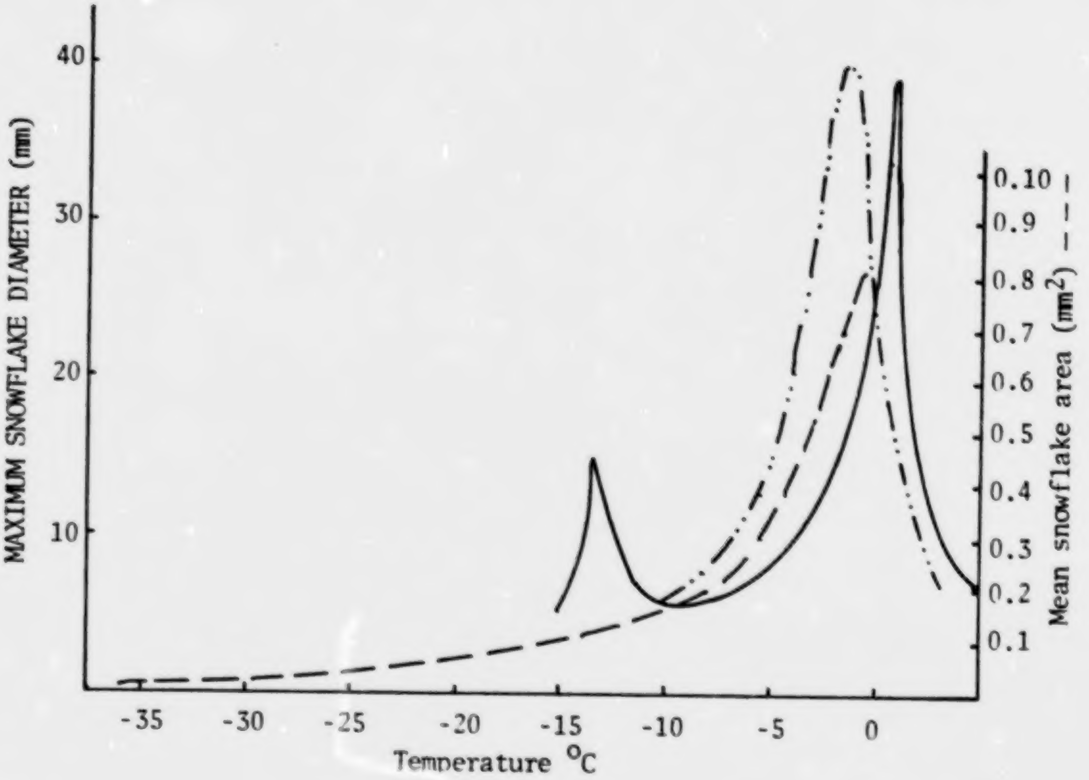


Figure 3.25 Size of snow crystal aggregates as a function of air temperature during snowfall (data from Spitzbergen, Westmann 1913: — — —; Japan, Magono 1953: —. .—; Wyoming, Rogers 1974a and b: ———). All curves fit by eye.

-10°C, and only 8% at temperatures less than -10°C. Magono (1953, 1954, 1960) found that snowflakes had their largest dimensions near -1°C, and that aggregation was mostly confined to temperatures warmer than -10°C. Hobbs et al. (1974a: Figure 8), working with frontal and orographic clouds over the Cascade Mountains, and Rodgers (1974a, b), who studied orographic cloud systems in Wyoming, confirmed these observations. The latter two studies also found a secondary peak in snowflake dimension centered between -12° and -17°C (Fig. 3.25).

The temperature dependence illustrated in Figure 3.25 reflects the net effect of many influences including ice particle concentration, sticking probability, and crystal type. Ice particle concentrations have a profound but joint effect with temperature on the probability of aggregation. Figure 3.26 illustrates smoothed isopleths of the probability of occurrence of aggregates as a function of air temperature and concentration of ice particles in the air. At temperatures below -15°C and for ice particle concentrations of less than 0.1 cm^{-3} aggregates are unlikely to form. For temperatures exceeding -5°C and particle concentrations in excess of 1 cm^{-3} there is more than a 50% chance of aggregate formation (Fig. 3.26). Heymsfield (1977) noted that ice particle concentration (crystals > 100 μm) increased with increasing air velocity and decreasing temperature. For values that are likely to be common on the coast (vertical velocity > 1 $cm \cdot s^{-1}$; temperature 0 to -20°C) the data of

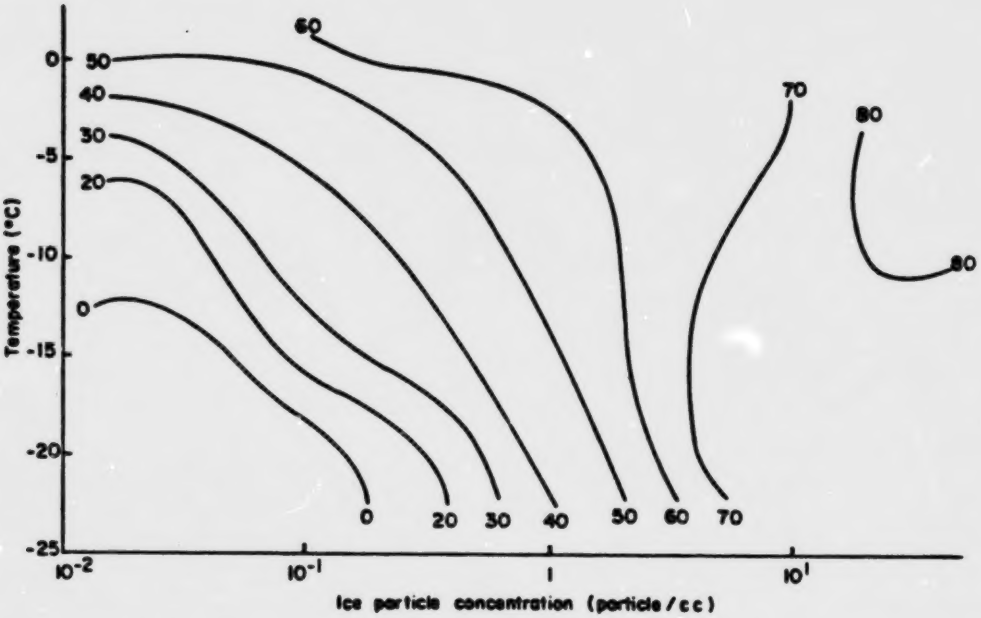


Figure 3.26 Isopleths for the probability (in percent) of finding ice crystal aggregates in a cloud as a function of the air temperature and the total concentration of ice particles in the air (from Hobbs et al. 1974a: 2203).

Heymsfield (1977: Fig. 6) suggest ice particle concentrations of 30 to >100 l^{-1} with most values greater than 100 l^{-1}. These data suggest that ice particle concentrations occasionally, but not commonly, could be limiting over coastal mountains (Figs. 3.18 and 3.26).

The effect of temperature is probably mediated through two phenomena. Laboratory observations show than on contact ice crystals 'stick' to each other by forming an ice bond across the surface of contact if the air temperature is close to 0°C. These adhesive forces decrease sharply as temperature declines below 0°C (Nakaya and Matsumoto 1954, Hosler et al. 1957). It is probably this phenomenon that produces the peak in aggregation near -1°C (Fig. 3.25).

Field observations of Hobbs et al. (1974b) and Jiusto and Weickmann (1973) demonstrate that most snowflakes are aggregates of planar snow crystals with dendritic features. However, aggregates of needles also occur (bundles, N1b, of Magono and Lee 1966, see Table 3.5). Aggregates of simple, thick ice plates and short columnar ice crystals are rare. Data of Hobbs et al. (1974a) from the Cascade Mountains show that aggregates of columns and needles tend to stay small, while aggregates of dendritic crystals tend to become large (Fig. 3.27). Although maximum snowflake diameters were as large as 15 mm, most snowflakes were 2 to 5 mm in diameter. These latter observations suggest that contacting ice crystals interlock with each other if the crystals have dendritic features. The interlocking mechanism would occur most

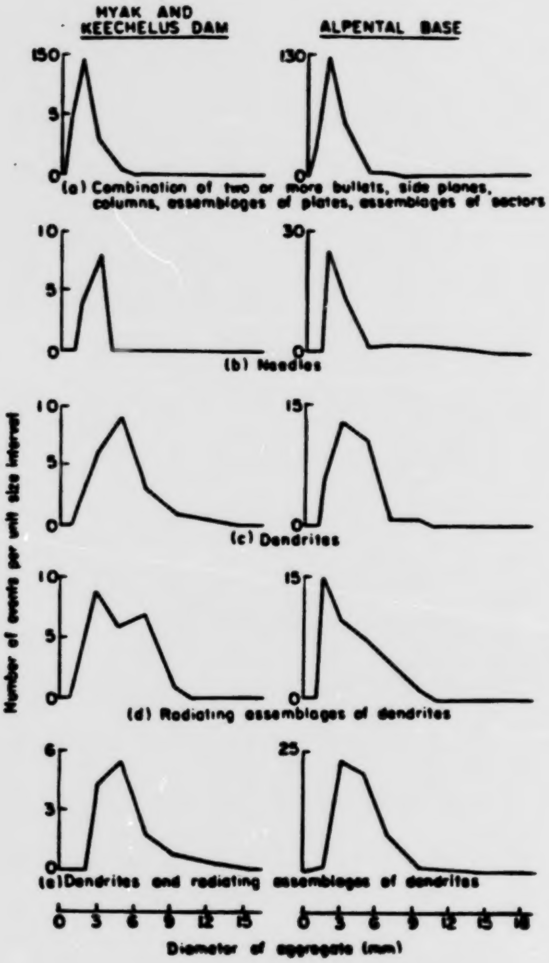


Figure 3.27 Size distributions of aggregates of different types of crystals collected at different sites in the Cascade Mountains (from Hobbs et al. 1974a: 2203).

frequently at temperatures between -12° and -17°C at relatively high saturations relative to ice because these conditions most favour dendritic features (Fig. 3.22). Interlocking probably explains the secondary peak with temperature (Fig. 3.25, also Hobbs et al. 1974a: Fig. 8).

Given the data on crystal shape from Mt. Seymour (Table 3.5) there are three implications. First, relatively warm coastal temperatures will encourage the formation of crystal aggregates (Fig. 3.25). Indeed, Hobbs et al. (1974a) noted that at ground level in the Cascade Mountains aggregates occurred with a frequency exceeding 60%. Second, because needle-shapes predominate, most assemblages will still be relatively small, about 3 to 4 mm in diameter (Table 3.5, Fig. 3.27). However, shapes of primarily dendritic form did occur in 29% or more of the snow storms on Mt. Seymour. Some of these (e.g., radiating assemblages of dendrites) probably have a mean diameter of about 6 to 9 mm (Fig. 3.27). Third, some crystal aggregates may have considerable mass. Snow crystal aggregates tend to follow predictable dimensional relationships during their clumping with other crystals. These relationships are expressible in terms of power laws of the form $M = aD$, where M is the snowflake mass, D is its maximum diameter, and a and b are constants for an aggregate of component crystals of given shape and bulk density.

Locatelli and Hobbs (1974) provided data relating dimensions to mass for 15 crystal types and crystal aggregates found over the Cascade Mountains (e.g., Fig. 3.28). Given an

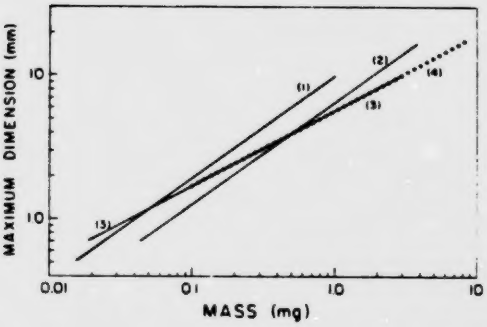


Figure 3.28 Best fit curves between mass and maximum dimensions for a number of crystal types from the Cascade Mountains, Washington (1) aggregates of unrimed side planes; (2) aggregates of unrimed, radiating assemblages of dendrites; (3) aggregates of unrimed, radiating assemblages of planes and bullets; (4) aggregates of densely rimed dendrites or radiating assemblages of dendrites (modified from Locatelli and Hobbs 1974: 2194).

expected maximum diameter exceeding 6 mm for radiating assemblages of dendrites (Fig. 3.27), and the observation that most particles were rimed, crystal aggregates probably averaged about 2 mg in mass. Individual crystals would be much smaller (Figs. 3.23b, 3.24b). Together these results suggest we should expect relatively high densities of new fallen snow in coastal mountains.

Growth by riming.--Riming is simply the accretion of water resulting from the collision of ice crystals with supercooled water drops. Snow crystals completely enveloped with rime are termed graupel or soft hail. For riming to occur both ice particles and supercooled water must exist. It is possible to compute theoretical collision efficiencies between ice particles and water droplets (Fig. 3.29a). The efficiencies computed refer to the collision of supercooled drops of radii between 1.0 and 55 μm with oblate spheroidal ice plates of different radii (147 to 404 μm). The most striking feature is the steep decline to zero efficiency at small and large drop sizes (this feature is in marked contrast to the behaviour of drop to drop collisions). The cutoff for drop radii of less than 10 μm agrees very closely with field observations of Hariyama (1975) for simple plate-like and dendritic crystals. Hariyama found that the number of droplets frozen on crystals decreased to 0.0 for radii near 10 μm (Fig. 3.29b). Note that field observations also corroborate the prediction of an upper size limit for collected drops.

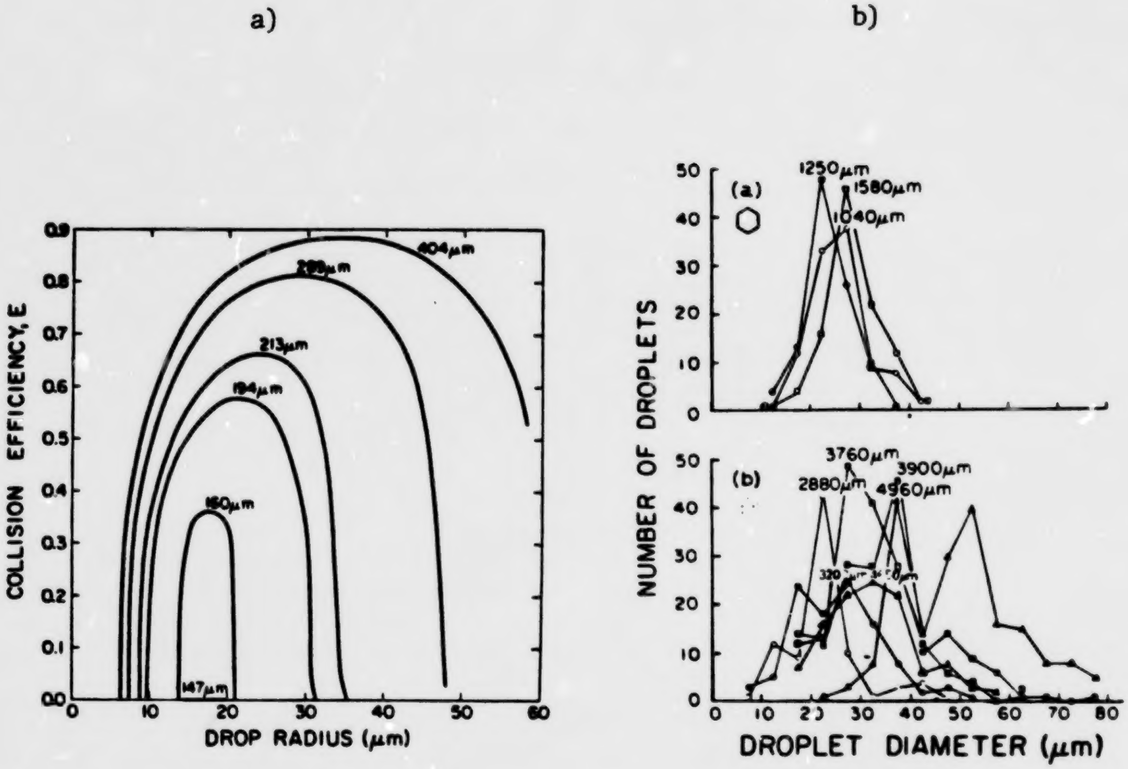


Figure 3.29 a) Theoretical efficiency with which thin, oblate spheroids of ice collide with supercooled water drops of various radii in air of -10°C and 700 mb. Numbers above curves refer to the dimension of the semi-major axis of the ice particles (from Pitter 1977: 685).
b) Observed size distribution of cloud drops accreted on planar snow crystals. Number labelling curve is the diameter of the collector crystal (from Pruppacher and Klett's (1978) adaptation of Hariyama 1975: 388).

Figure 3.29a also suggests a minimal crystal size below which drops cannot be collected. Theoretical calculations predict the limiting radius to be about 150 µm which is in excellent agreement with observations of Ono (1969), Wilkins and Auer (1970), and Hariyama (1975). They found that simple hexagonal ice plates must grow by vapour diffusion to a size greater than 150 µm radius before they can commence riming (e.g., Fig. 3.30). Note that Figure 3.30 also shows that the onset and intensity of riming shifts to progressively larger sizes with increasing dendritic shape of the planar snow crystal. The latter result is expected because the fall velocity of a crystal of given size decreases with increasing dendritic features (Fig. 4.7b). Theoretical computations of Schlamp and Pruppacher (1977) predicted that radii for needles (the most common crystal type on Mt. Seymour) must attain 15 to 20 µm before onset of riming, in good agreement with field observations of Reinking (1976).

These collision efficiencies have several implications to coastal snowfall. The data presented suggest that planar and needle crystals must grow by diffusion to diameters of 300 and 30-40 µm, respectively, before they can commence riming. We see from Figure 3.24b that a P1c crystal growing by diffusion at water saturation and -15°C requires less than a minute before riming is possible. Given the dimensional relationship illustrated (Fig. 3.23b) needles also require less than a minute at -6° to -7°C before onset of riming. Providing ice-forming nuclei and liquid water are present there is

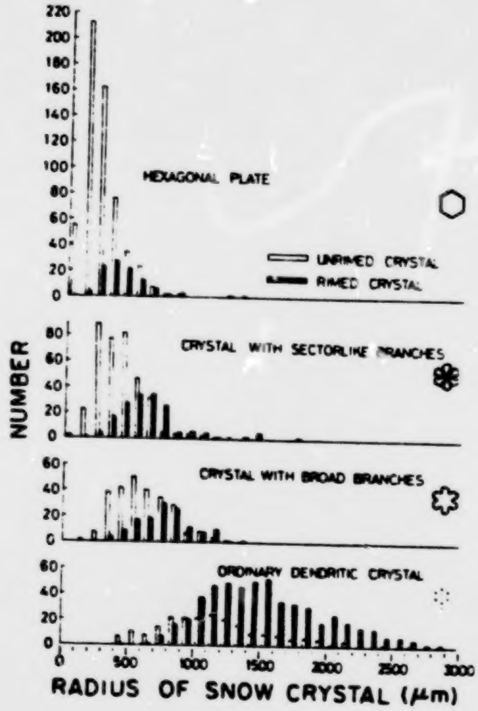


Figure 3.30 Observed relationship between the onset of riming and the radius of planar snow crystals. Open columns represent unrimed crystals; solid columns represent rimed crystals (from Hariyama 1975: 385).

obviously ample time for orographic cooling, ice crystal formation, and riming to occur frequently on the coast. Furthermore, given the probable size distribution of drops within the clouds (Fig. 3.19a), and the apparent collision efficiencies of drops of that size (Figs. 3.19b and 3.29a), riming should proceed rapidly. Together, these observations suggest that most snowfalls on the coast will involve particles of relatively high moisture content and bulk density.

Graupel occurred in 62% of the coastal snow storms sampled (Table 3.5). Hariyama (1976) carefully sectioned and disassembled natural graupel particles under a microscope to find both snow crystals and frozen drops as central particles. Whatever the central particle, if it grows by collision with supercooled drops, it will have a surface temperature higher than that of the surrounding air owing to the release of latent heat during freezing. As long as the latent heat of freezing is dissipated from the growing ice particle such that its temperature remains below 0°C, all accreted cloud water must freeze to the ice particle. The particle is then considered to grow in the 'dry growth regime' sensu Ludlam (1958). With increasing liquid water content of the cloud, increasing drop size, increasing frequency of collision between drops and the ice particle, the temperature of the growing particle gradually rises (the rate of temperature increase decreases as the surface temperature of the particle approaches 0°C). Under such growth conditions ('wet growth regime' sensu Ludlam 1958) not all accreted water is converted to ice and the amount of ice formed is determined by the rate at which heat is dissipated.

Conditions in the coastal mountains would encourage the wet growth regime. Initial studies of the wet growth regime assumed that the growing ice particle would remain solid and shed all excess water (Schumann 1938, Ludlam 1958); that assumption has since been invalidated by wind tunnel experiments (List 1959, 1960; Macklin 1961). Current theory exceeds the ability of researchers to document the required parameters under natural conditions. Moreover, the processes are stochastic, particularly when the ice particles and water droplets are of similar size. In short, there are few ways to connect theoretical microphysical treatments of graupel formation with the macromeasurements acquired in nature. Two broad implications are evident. First, given the growth of graupel by accretion of water rather than by aggregation or clumping of crystals, bulk densities of individual graupel particles should be greater than for snow flakes. Second, the resultant bulk densities should be a function of freezing rates. At slower freezing rates more dissolved air can escape by diffusion and more bubbles can migrate to the surface of the accreting ice particle. Slow freezing rates lead to relatively clear ice, and rapid freezing rates lead to relatively opaque ice. Carras and Macklin (1975) found bubble concentrations of 10^6 to 10^8 cm^{-3} typical of dry growth regimes with rapid freezing rates and 10^5 to 10^6 cm^{-3} in wet growth regimes with relatively slower freezing rates. Given these observations and the meteorological conditions under which graupel forms in south coastal British Columbia, it follows that much of the snowfall should be relatively dense.

3.4 Summary - Processes Creating Snow

Ice crystals form when condensation nuclei are present in moist air at temperatures at or below freezing (Fig. 3.12). When there is sufficient time for the crystals to grow to measurable mass and fall, snow is delivered to the ground. Laboratory experiments and empirical observations indicate that once ice crystals form under the conditions prevailing in south coastal British Columbia, rapid growth ensues (Fig. 3.24, Ch. 3.3.2). Time for ice crystals to grow is unlikely to be limiting there. The pattern of abundance of cloud condensation nuclei appears erratic (Fig. 3.15) and their concentrations may occasionally be limiting to snowfall. It is more likely, however, that temperature and moisture limit snow delivery to the ground. We attribute the variability in local ice crystal concentrations (Fig. 3.18) largely to variability in the relative humidity of the air mass (Ch. 3.1.1). However, moisture and temperature are covariate in air masses, and both exhibit considerable range in south coastal British Columbia.

The unpredictable nature of snowfall in southern British Columbia appears to result from two broad features: the variable nature of air masses present and the pronounced orographic effect. Although dominated by the Aleutian Low during winter, the large scale weather patterns encompass a variety of air masses (Fig. 3.4). Furthermore, these masses approach the coast following different trajectories (Fig. 3.6) which produce different moisture contents. Generally, the moisture content is high (Fig. 3.5). The cyclonic nature of the large weather systems encourages convergence of different air masses producing frontal conditions. About 83% of the storms on Mt. Seymour were frontal (Table 3.2). Because the weather system can attract a variety of air masses, the nature of the front is changeable (Tables 3.1 and 3.2). The high moisture contents usually result in precipitation, but temperatures may not be sufficiently cold to produce snow. Conditional probabilities of snow computed for different storm types on Mt. Seymour (Table 3.2) represent first approximations, but may not be general. Furthermore, the Aleutian Low can shift between years thus generating different storm patterns.

The strong orographic effect also increases the difficulty in predicting snowfall because it introduces a strong effect of elevation. Conditions in southcoastal British Columbia encourage orography, but the specific effect varies with the storm type involved (Table 3.3) and likely the local relief. Data reviewed indicate that the effect is broadly predictable on a local scale, providing the freezing level is known (Eqs. 3.2 and 3.3). The freezing level itself is a function of air mass and storm type (Table 3.4). Because the lapse rate appears stable (Ch. 3.2), snowfall should increase monotonically with elevation in southcoastal British Columbia. Given the empirical lapse rate of $7^{o}C \cdot km^{-1}$, the capacity of air to retain moisture (Fig. 3.1), and the nature of coastal air masses (Ch. 3.1), most winter storms in southcoastal British Columbia should produce snow above 1000 m elevation. Particular air masses and storm types can produce a freezing level sufficiently high that no snow falls regardless of the orographic effect (Table 3.4), but these storm types are relatively rare (Table 3.1). An obvious deduction is that snowfall at lower elevations must be much more variable and less predictable than at higher elevations. It is apparent why most game animals traditionally move to lower elevations. There are two generalities relevant to management: i) predictability of snowfall on lower elevation winter ranges will always be poor; and ii) unless a strong warming trend occurs, lower elevation winter ranges will remain important, as a result of strong and locally predictable orographic effects.

The kind of snow that will fall is much more predictable than whether snowfall will occur (Fig. 3.21 and 3.22). At elevations of ungulate winter ranges, aggregates (snow flakes) will be common (Fig. 3.25 - 3.27), but most assemblages will be small because needle shapes will predominate (Fig. 3.22). Many aggregates will have considerable mass (Fig. 3.28). That mass coupled with a high probability of riming (Ch. 3.3.2, Table 3.5) means that newly fallen snow will be relatively dense. Generally, it will be denser that the value of 0.1 g cm^{-3} commonly assumed. The increased density will decrease ungulate sinking depth, but increase drag during locomotion.

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4. SLOPE AND WIND

Discussion within this chapter remains restricted to abiotic features. Two are treated, slope and wind. Discussion of the interaction of these features with biotic components, such as forest cover, is deferred. For example, the slope or angle of orientation of tree branches influences interception of snow but that phenomenon is addressed during discussion of interception by individual trees (Ch. 7.6.2). Similarly, wind modifies apparent interception by stands which is discussed in Ch. 6 and 8. Here we restrict discussion to delivery of snow to an intercepting surface. Redistribution of fallen snow is not treated; however, the mechanics of wind and their relationship to such features as surface roughness are discussed. These mechanics influence both delivery and redistribution of snow. Only general principals are noted. The influence of slope on the formation of precipitation (orographic effect) has been discussed (Ch. 3.2.1).

4.1 Slope and Snow Delivery

Although many studies have examined potential relationships between slope and snowpack, the effect of slope on snow delivery has received minimal treatment in literature known to us. Most literature treats slope in terms of its effect on the snowpack through its influence on melt and mass movement (creep and avalanches). In terms of snow delivery the most obvious theoretical influence of slope is the increase in effective ground surface relative to horizontal area with increasing slope. Theoretically, snow delivery per unit area on a slope should be related to snow delivery on the level by the cosine of the angle of the slope (Fig. 4.1). If snow delivery was vertical, a 30° slope would receive 87% as much snow per unit area as level ground, whereas a 60% slope would receive only 50%. The relationship should also apply to individual branches within a tree.

Differences in snow delivery of this magnitude (13 to 50%) could be significant, particularly if cumulative with each storm. Lower amounts of snow on steeper slopes frequently are attributed to greater wind speeds or greater solar angles, but also may be a product of the simple geometric relationship noted (Fig. 4.1).

Were the simple geometric effect acting, it would increase the potential for winds of a given velocity to redistribute snow on steeper slopes. On steeper slopes the sparser distribution of incoming snow would reduce adhesive and cohesive forces. The potential confounding of solar angle could be eliminated by treating only the accumulation phase, but other factors remain. The relationship of Figure 4.1 assumes snow is falling vertically. Vertical descent may be approximated by wet heavy snow, but is unlikely to be true of drier snow. The orientation of the snow-forming surface to the ground also would interact with the rate at which particles were produced and contacted the ground surface

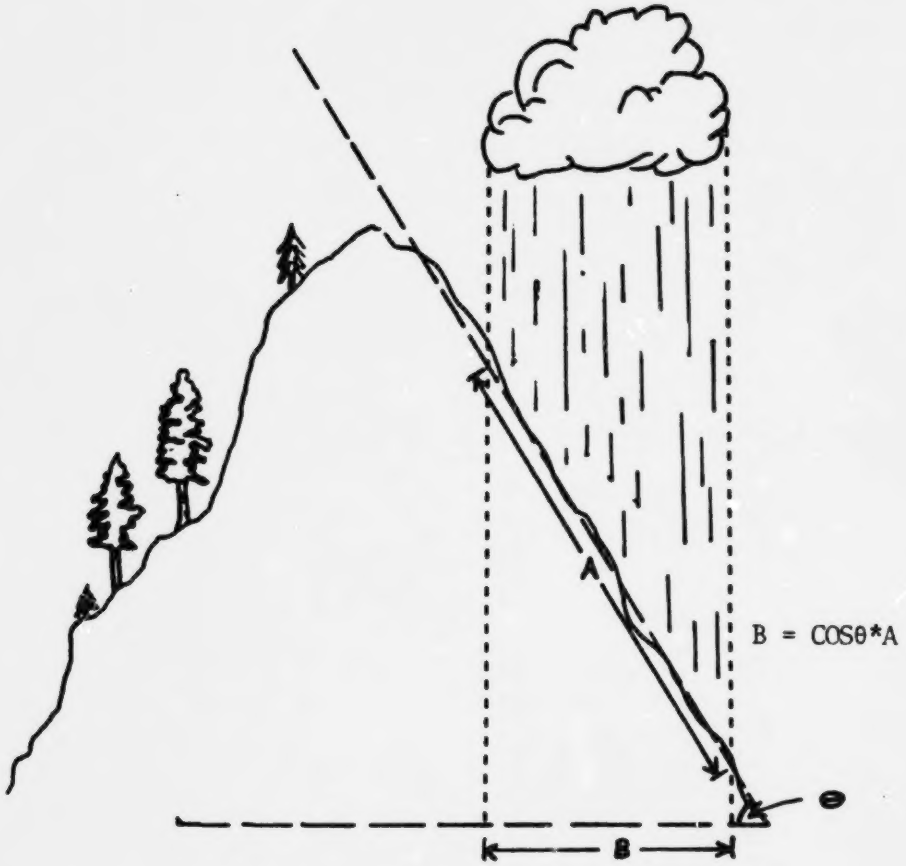

Figure 4.1 Potential effect of slope upon the density (snow crystals·unit area^{-1}) of snow delivery to the ground.

(Fig. 4.2). That rate could itself be modified by wind (note different 'angles of attack' of snowflakes in Figs. 4.2a and b). Because snowfall in coastal areas is often associated with rapid rates of orographic lifting, the effects of Figure 4.1 could be countered by those illustrated in Figure 4.2b. Actual airflow is complex and 3-dimensional; a more detailed treatment of the interaction of slope and wind velocity is presented in Ch. 7.6.2. That treatment extends the predictions of Figures 4.1 and 4.2 to smaller surfaces and evaluates their accuracy.

4.2 Wind and Snow Delivery

The effects of wind or air motion occur at vastly different scales, from movement of air masses (Fig. 3.3) through redistribution of snow during a storm (Fig. 6.5), to interception within individual trees (Fig. 7.7 and 7.10b). Here we concentrate on snow delivery to the ground regardless of type or pattern of vegetation present, and briefly summarize processes influencing wind or air movement.

4.2.1 Processes Influencing Wind

Wind type.--The general circulation of the atmosphere is driven ultimately by the imbalance of solar radiation received over the earth's surface and modified by the earth's rotation. Within, and contributing to, the general circulation are zones

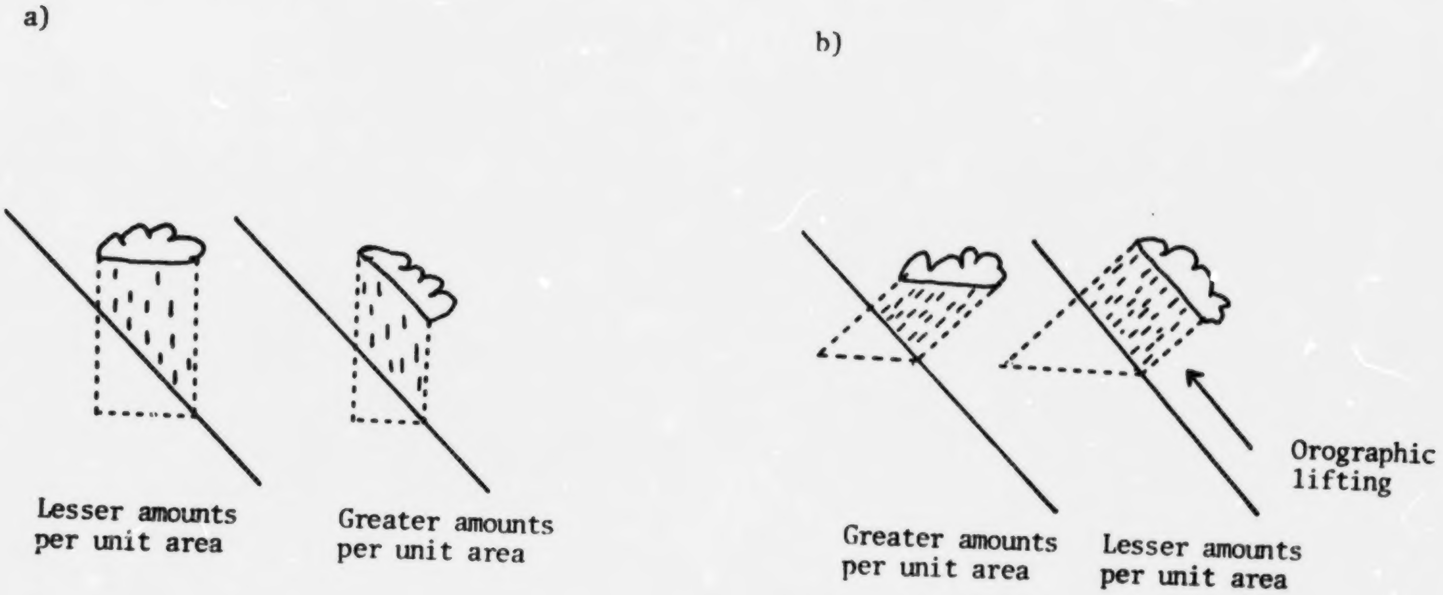
a)
b)
Lesser amounts per unit area
Greater amounts per unit area
Greater amounts per unit area
Lesser amounts per unit area
Orographic lifting

Figure 4.2 Greatly simplified representation of the interactions between slope, angle of the snow-forming surface, and rates of snow production.
a) Low rates of snow production.
b) Rapid rates of snow production.

of high and low tressure. Air masses themselves generally form over relatively homogeneous regions often characterized by high atmospheric pressure (Ch. 3.1). The discontinuities or frontal zones between air masses are areas of low armospheric pressure. Differences in pressure create gradients along which air flows, or wind blows. At a practical scale of regional winds, the fundamental wind type above the boundary layer (in the 'free' atmosphere) is geostrophic. At these altitudes (greater than about 300 to 1000 m above the ground) the motion of the air (wind) is such that the pressure-gradient forces are approximately in equilibrium with Coriolis forces. These geostrophic winds are governed by the pressure distributions associated with large-scale weather systems and their characteristics are independent of the surface topography. The horizontal air flow can be closely approximated by the geostrophic wind speed Ug, where:

$$Ug = \alpha\ (\Delta P/\Delta N)/f \qquad (4.1)$$

and α = specific volume,
ΔP = horizontal pressure gradient,
ΔN = horizontal perpendicular distance over which ΔP is measured, and
f = Coriolis parameter, a constant for a given latitude (we are indebted to D.G. Schaefer pers. commun. for this modification and correction of Berry 1981).

The broad patterns of air movement along B.C.'s coast were reviewed in Ch. 3.1. The geostrophic wind blows perpendicular to the horizontal pressure gradient at a speed proportional to the gradient, with the low pressure to the left of an observer facing downstream in the northern hemisphere. Coriolis force is an apparent rather than a real force introduced to account for the rotation of the coordinate system (which is fixed to the rotating earth). If earth did not rotate, winds would simply flow from high to low pressure and meteorology would be much less interesting. Deflection of winds to the right (in the northern hemisphere) results in geostrophic flows parallel to, rather than directly across, isobars.

Although regional winds along B.C.'s coast are generally and approximately geostrophic, substantial ageostrophic effects are important. For example, when a low pressure storm centre is deepening rapidly, as they often do in this area, "isallobaric" winds flow across isobars toward the low. Also, in mountainous areas, topographic barriers extend well above what is normally considered to be the boundary layer, thereby influencing winds in what is normally the geostrophic layer. These influences include local orographic lifting.

"Local winds" may deviate from those of the large scale circulation pattern. Winds at this scale are generally at lower elevations and influenced more by local topography. The topography, especially in the western mountainous regions of North America, can cause channeling and compaction of air flow increasing the air pressure within a small area. Local winds can also be driven by thermal effects (land and sea breezes, mountain and valley winds, chinooks, etc.). Again, movement occurs from high to low pressure areas.

Surface roughness and the velocity profile.--At the earth's surface the so-called "no-slip condition" requires that the wind speed be zero (Kind 1981). Therefore there must be some "boundary" or "friction" layer through which the speed increases from zero at the earth's surface to its geostrophic value at the top of the layer. The thickness of the layer varies from about 300 to 1500 m (Geiger 1961, Berry 1981, Kind 1981). Geostrophic or true gradient winds exist only above the friction layer. By definition the friction layer influences both local and general winds in terms of wind velocity, wind direction, and flow patterns. Both the thickness of the layer and the distribution of wind velocity with height Z above ground level (the "velocity profile") depend on surface roughness. Over relatively rough terrain the boundary layer is thicker and wind speed increases relatively slowly with height; over flat, open terrain the opposite is true (Fig. 4.3).

The wind speed quoted in weather reports is normally measured at a height of 10 m at a local airport site where terrain is flat and open; it generally differs from the wind speeds at other heights and/or over different terrain.

The relationship between wind velocity at the gradient level (Ug) and that at a lower level, down to the lowest few

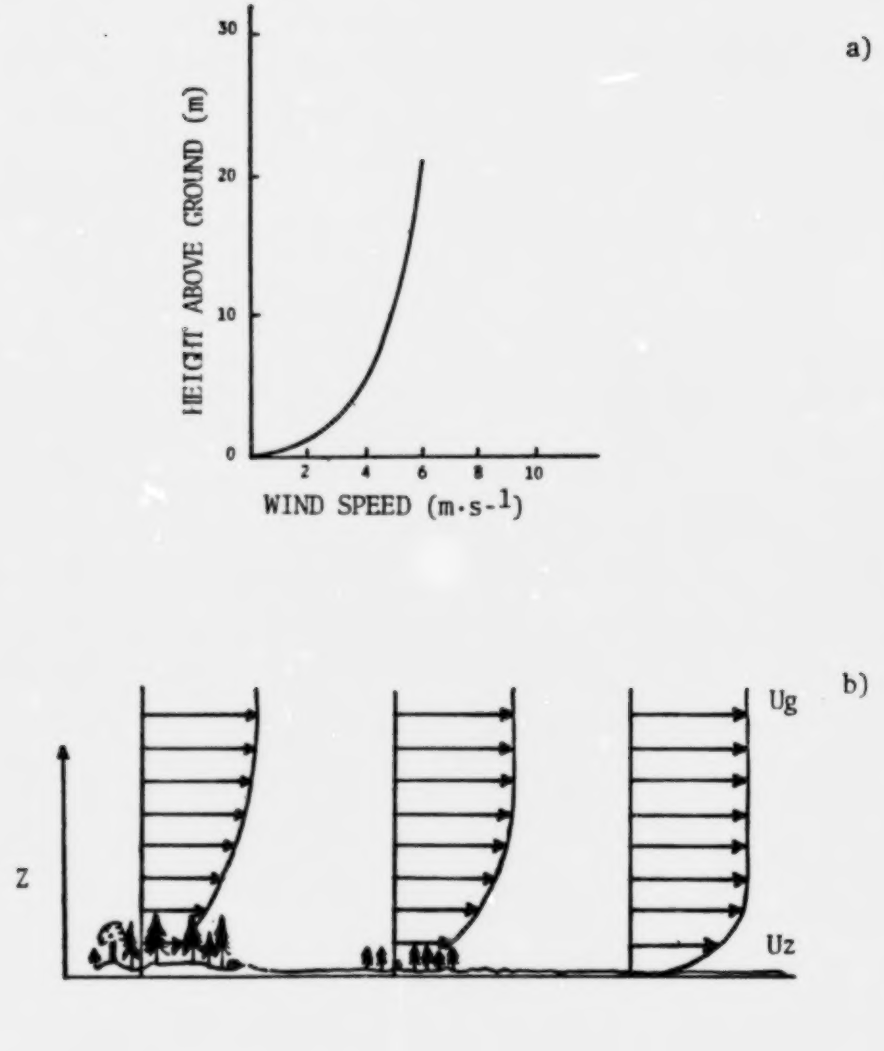

Figure 4.3 a) Measured increase in wind speed with height above ground in open terrain (modified from Miller and Thompson 1975: 93).
b) Schematic representation of velocity profiles over terrain with different roughness (modified fron Kind 1981: 339).

metres above the ground is usually well represented by the power law (e.g., Fig. 4.3):

$$Uz/Ug = (z/Zg) \qquad (4.2)$$

where Uz is wind velocity at height z, Zg is height of the geostrophic wind, and n is an exponent whose magnitude depends on the roughness of the underlying surface and the stability of the air (indicated by the vertical temperature gradient; neutral conditions are approximately $-10°C \cdot km^{-1}$).

Values for n are somewhat site specific and are related to stability of the air. DesMarrais (1959) using measurements taken over rolling, partially forested terrain calculated values for n ranging from 0.19 for superadiabitic conditions ($dT/dZ < -11°C \cdot km^{-1}$) to 0.49 for inversion conditions ($dT/dZ > 0$). Under strong inversions over an airfield ($dT/dZ > 16°C \cdot km^{-1}$), Frost (1948) reported an average value for n equal to 0.77. These values illustrate the effects of stability on wind speed at a given height relative to that at the surface; the more unstable the air mass, the larger the Uz/Ug ratio. Because of its insulative and radiative properties snow cover tends to increase the stability of the adjacent atmosphere and moderates the change in velocity profiles.

Winds near the ground are affected more by the roughness of the underlying surface than stability of the air (Fig. 4.3b); the value of n in Eq. 4.2 is influenced by the frictional drag exerted by the surface. Surface roughness is often quantified in terms of the roughness coefficient or parameter of Prandtl (1957). The parameter or height Zo varies with the average height of obstacles on an otherwise smooth surface and allows a numerical value to be assigned to the aerodynamic roughness of the underlying surface. The dependency of wind speed on surface roughness is difficult to determine, mainly because it has a complex relationship with stability. Under neutral conditions (dT/dZ about equal to $-10^{\circ}\mathrm{C}\cdot\mathrm{km}^{-1}$) the wind velocity profile over relatively smooth surfaces, up to a height of about 30 m, can be represented by the logarithmic profile:

$$Uz = U^*/K \ln(Z/Zo) \tag{4.3}$$

where Uz = wind velocity at a height, Z, above the surface,

U* = friction velocity(= τ/ρ where τ is the shear stress and ρ is the air density,

Zo = roughness height, and

k = von Karman's constant (usually taken as 0.4)

Some representative values of the roughness height of different surfaces provided by Deacon (1953), Baumgartner (1956), Byers (1959: 515), Satterlund and Eschner (1965), and Munn (1966) are:

Surface	Zo (cm)
Smooth mud flats	0.001
Snow	0.001 to 0.07
Smooth snow on short grass	0.005
Short grass (1.5 to 3 cm)	0.02 to 0.7
Long grass (60-70 cm)	3.7 to 9.0
65-year-old Scots pine and 50-year-old dense Ponderosa pine	30 to 175
Park-like distributions of trees	100
Young dense pine plantation	290

The effect of snow cover on surface roughness and the wind velocity-height profile can be demonstrated by Eq. 4.3. When snow covers short grass, Zo is reduced from about 0.4 to 0.005 and the velocity profile is altered such that wind speeds are greater with the differences becoming larger with decreasing height (Fig. 4.4)

Minimum wind speeds necessary for turbulent flow also vary with the roughness coefficient or underlying surface (Fig. 4.5). Natural surfaces vary markedly and only the relative value of Zo is reliable. At one end of the scale, a smooth snow cover over short grass has a relatively low roughness parameter because strong winds are needed before turbulent motion above that surface is attained (Figs. 4.4 and 4.5). Conversely, forests can cause turbulent motion at very low wind speeds and thus have a relatively high roughness

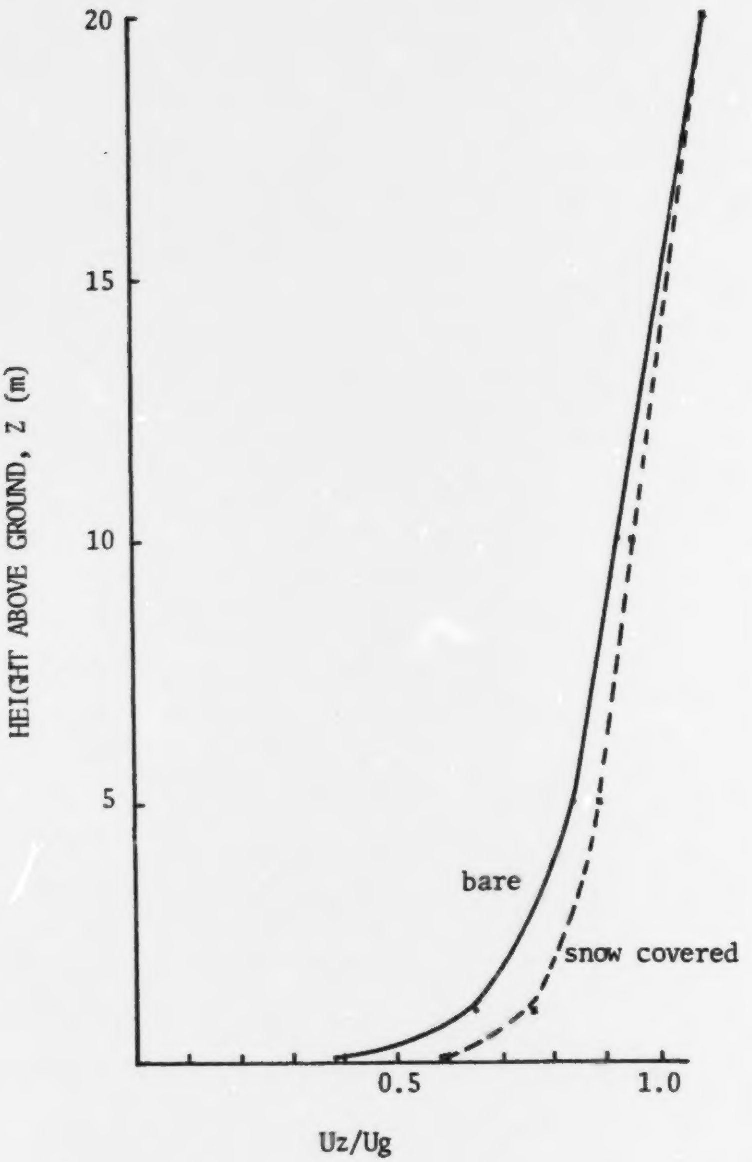

Figure 4.4 Effect of snow cover over short grass on the wind velocity profile (see text for calculation).

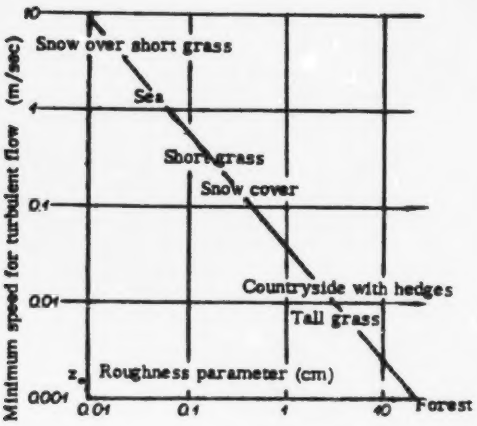

Figure 4.5 Roughness coefficients for different natural surfaces (modified from Geiger 1961: 275 after Deacon 1953).

coefficient (Fig. 4.5).

In the aerodynamic sense, the surface roughness of an obstacle depends on its height, breadth, and porosity or penetrability (Geiger 1961). Mountains represent the maximum degree of surface roughness thereby providing the greatest friction to air flow. Generally, prevailing winds blow at right angles to mountain ridges (Yoshino 1975). Warm air is deflected up over the mountains increasing in velocity as it goes; cold air is mostly reflected. As air passes over a ridge it reacts according to the ridge shape and orientation. Round topped ridges cause little turbulence while bluffs or sharp ridges always create some degree of turbulence (U.S. Dept. of Agric. 1964). Large roll eddies are typical of sharp ridge tops or bluffs (Fig. 4.6). Turbulence caused by a mountain range increases as wind velocity increases, and as the angle of prevailing wind approaches a perpendicular flow to the mountain range.

4.2.2 Effects of Wind

Wind has obvious influences on snow delivery, but influences which are difficult to quantify and to generalize. Three broad effects are possible, each with its own implications: i) contribution to orographic effects; ii) modification of ice crystals; and iii) diagonal vectors to snowfall.

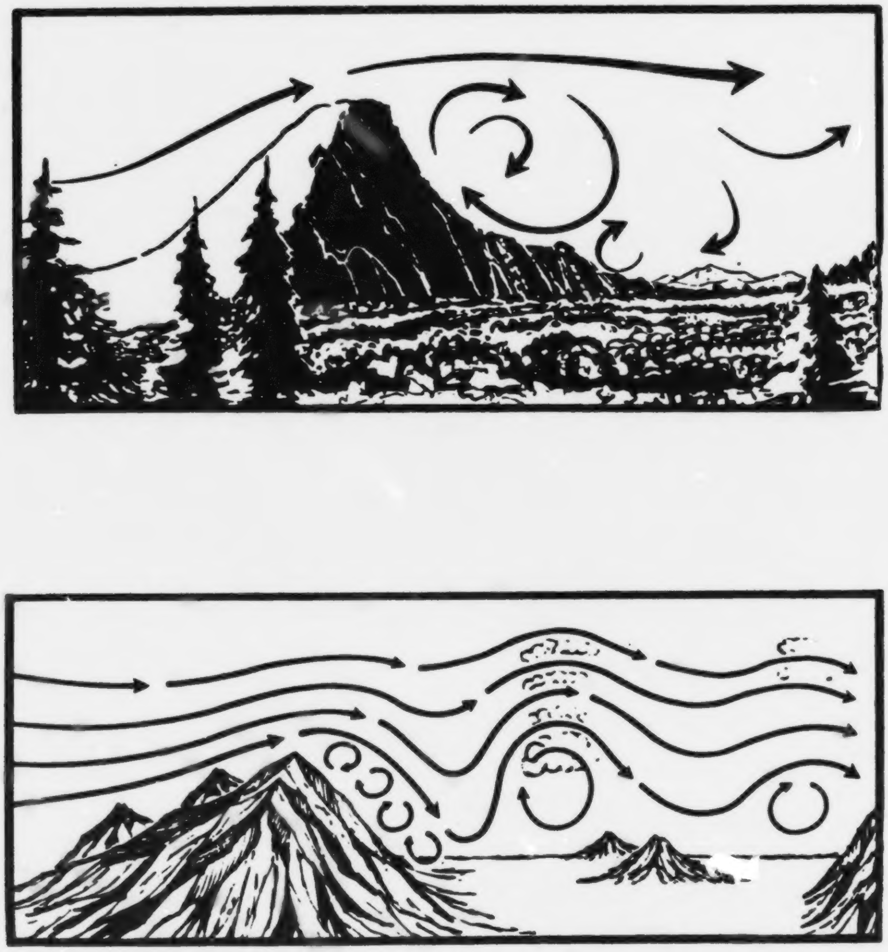

Figure 4.6 Schematic representation of roll eddy produced on the leeward side of a ridge (from USDA Forest Service 1964: 16-17).

Orographic effect.--Orographic effects have been discussed (Ch. 3.2.1). In general, higher wind speeds cause greater rates of precipitation at higher elevations. Fulkes (1935) derived an approximate relationship for the precipitation rate P of an ascending layer of saturated air of unit cross section:

$$P = bW\Delta Z \tag{4.4}$$

where b = a coefficient whose value depends on the temperature and pressure of the layer (height),
W = vertical air velocity, and
ΔZ = thickness of the layer.

Under conditions when most snow forms (temperatures below 0°C and heights below 6 km) the value of b decreases rapidly with decreasing temperature and, to a much lesser extent, with decreasing pressure (increasing height). In coastal British Columbia temperatures are seldom very low, nor are heights very high. Thus the potential precipitation rates are high and directly proportional to vertical velocity (Eq. 4.4). The vertical velocity is determined mainly by the characteristics of individual weather systems and the extent to which terrain features affect air flow, for example, by orographic lift. These features were reviewed in Chapters 3.1 and 3.2, and it was noted that conditions favoured rapid orographic lifting.

Ranahan and Alexander (1979) noted another positive effect of wind velocity on precipitation rate. They suggested that high winds are necessary to produce sufficient turbulence to break up low level inversions, which in turn increases the likelihood of snowfall by encouraging a near neutral lapse rate. However, in some cases low level inversion is necessary to maintain temperatures cool enough for snow at the surface.

Snow crystal structure.--Effects of wind on snow crystal structure are less clear. It is well documented that strong updraughts (convective winds) are important in the formation of large hail stones, because updraughts prolong the period of riming or accretion. It is also possible that slight turbulence encourages the growth of snow crystal aggregations by enhancing the probability of contact and aggregation. However, the largest snowflakes apparently are formed at very low wind velocities, and the more probable effect of wind on snow crystals is to mutilate or deform them. Individual crystals falling on windless days when temperatures are near freezing may merge to form massive snowflakes up to 15 mm in diameter.

At greater wind speeds snowflakes may be broken down into their elemental, needle-shaped components. This effect of wind would be more evident in the character of snow delivered, than in the delivery pattern. Evidence that the wind has strong influences on the density of newly fallen snow on steep slopes presented in Ch. 7.6.2. Wind modification of ice-crystal structure could alter the effectiveness of adhesive and cohesive forces and influence the delivery pattern by affecting redistribution during snowfall. Such effects are documented in Table 7.1 and Fig. 7.10.

Diagonal delivery.--The third and most obvious effect of wind on snow delivery is that wind causes snow to fall diagonally. Snow arrives from all directions and is transported by eddies and other turbulent air motion. Anderson (1970) observed that snow does not simply fall "...it swirls about until it finds a surface to attach to ...only to take off again to the atmosphere or to a new ground location." Rikhter (1945) observed that rarely does snow remain at the spot where it fell. This effect will likely be less pronounced along B.C.'s coast where snow is generally denser and wetter than elsewhere.

We lack confidence in predicting where snow eventually will accumulate because of the huge variation in air flow (Figs. 4.3 and 4.6; see also Ch. 8). The basic processes involved during wind transportation of snow include the characteristics of air motion noted above (Ch. 4.2.1), the characteristics of snow and processes of adhesion and cohesion (Ch. 3.3 and 7.1), and the characteristics of intercepting surfaces (Ch. 7.6). Here we discuss the manner in which different kinds of snow fall or are delivered to the ground.

The influence of wind on falling snow is pronounced because of the low bulk densities and high drag coefficients of many ice crystals. Terminal fall velocity of raindrops increases almost linearly with drop radius; the velocity of most snow crystals increases more gradually (Fig. 4.7). Note also that the terminal velocities of ice crystals are only 10 to 15% as rapid as those of raindrops (Fig. 4.7). For unrimed crystals, most crystal aggregates fall with speeds of 1.0 to 1.5 $cm \cdot s^{-1}$, which increase very little with increasing size of the aggregate (Fig. 4.8).

Rimed crystals and graupel fall more rapidly. Locatelli and Hobbs (1974) provided data from areas 750 to 1500 m above sea level in the Cascade Mountains of Washington (Table 4.1). From their data and those of Kajikawa (1972), we can extract six conclusions: i) terminal velocity increases as maximum dimension of the particle increases; ii) terminal velocity increases as the mass of the ice particle increases; iii) terminal velocity of a densely rimed ice particle may be up to twice as large as that of a similar unrimed ice particle of the same maximum dimension; iv) for a given maximum dimension terminal velocity of a particle increases with increasing density; v) terminal velocities of aggregates are generally greater than those of their component crystals (c.f. Fig. 4.7b and Fig. 4.9a); and vi) significant differences exist in the terminal velocities of different types of particles even when the particles have similar degrees of riming (Fig. 4.8b).

Kajikawa (1972, 1975a and b) provided formulae similar to those of Locatelli and Hobbs (1974), but utilized melted diameter rather than crystal dimension. His data nevertheless yield similar conclusions.

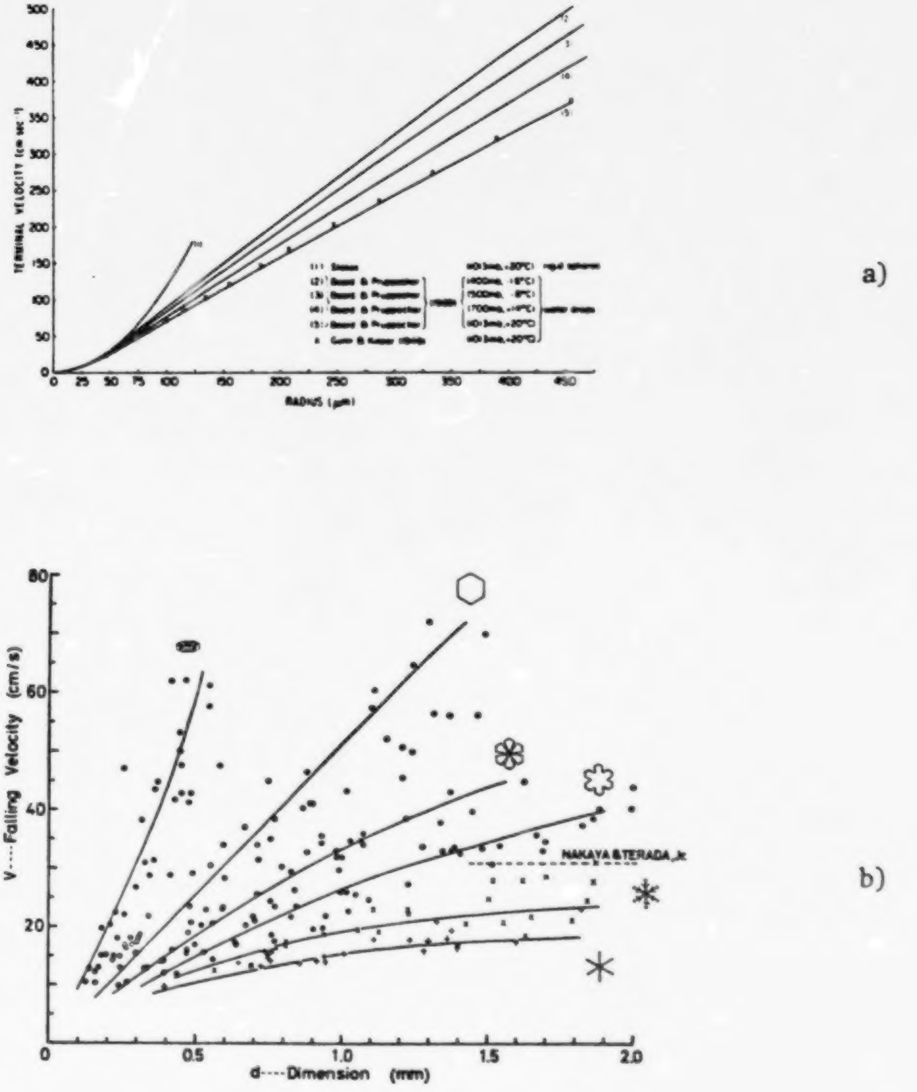

Figure 4.7 a) Terminal fall velocity as a function of size of water drops smaller than 500 µm in air (from Beard and Pruppacher 1969: 1071).
b) Terminal fall velocity as a function of dimension of planar type snow crystals (from Kajikawa 1972: 582).

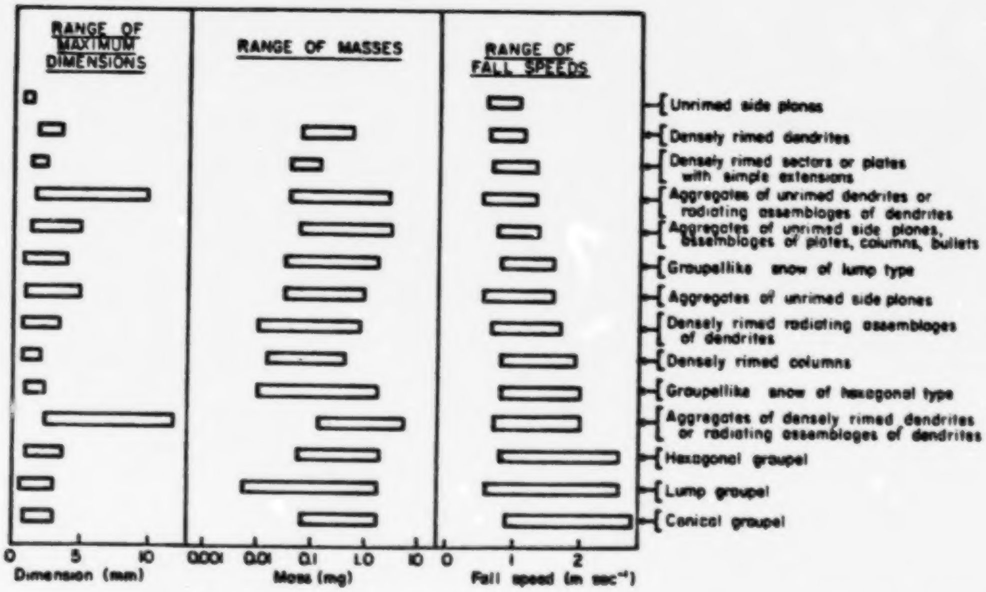

Figure 4.8 Ranges of maximum dimensions, masses, and fall velocities of solid precipitation particles over the Cascade Mountains (from Locatelli and Hobbs 1974: 2187).

Table 4.1 Derived relationships between fall velocities, masses and maximum dimensions for solid precipitation particles over the Cascade Mountains (from Locatelli and Hobbs 1974: 2188).

Particle Type	Velocity-Size Relationship	Velocity-Mass Relationship	Mass-Size Relationship	Density Range, Mg m^{-3}	Range of Maximum Dimensions, mm
Lump graupel	$V = 1.16D^{0.46}$, $N = 35$, $r = 0.55$	$V = 1.3M^{0.15}$, $N = 35$, $r = 0.53$	$M = 0.042D^{3.0}$ $N = 35$, $r = 0.98$	0.05 to 0.1	0.5 to 2.0
Lump graupel	$V = 1.3D^{0.66}$, $N = 58$, $r = 0.77$	$V = 2.4M^{0.24}$, $N = 58$, $r = 0.84$	$M = 0.078D^{2.8}$, $N = 58$, $r = 0.93$	>0.1 to 0.20	0.5 to 3.0
Lump graupel	$V = 1.5D^{0.37}$, $N = 17$, $r = 0.58$	$V = 1.8M^{0.12}$, $N = 17$, $r = 0.52$	$M = 0.14D^{2.7}$, $N = 17$, $r = 0.98$	>0.2 to 0.45	0.5 to 1.0
Conical graupel	$V = 1.2D^{0.65}$, $N = 30$, $r = 0.70$	$V = 2.5M^{0.28}$, $N = 26$, $r = 0.81$	$M = 0.073D^{2.6}$, $N = 26$, $r = 0.91$	...	0.8 to 3.0
Hexagonal graupel	$V = 1.1D^{0.57}$, $N = 33$, $r = 0.77$	$V = 2.0M^{0.18}$, $N = 31$, $r = 0.76$	$M = 0.044D^{2.9}$, $N = 31$, $r = 0.93$	...	0.8 to 3.2
Graupellike snow of lump type*	$V = 1.1D^{0.28}$, $N = 17$, $r = 0.46$	$V = 1.4M^{0.08}$, $N = 17$, $r = 0.32$	$M = 0.059D^{2.1}$, $N = 17$, $r = 0.91$	...	0.5 to 2.2
Graupellike snow of hexagonal type†	$V = 0.86D^{0.25}$, $N = 22$, $r = 0.38$	$V = 1.4M^{0.14}$, $N = 22$, $r = 0.71$	$M = 0.021D^{2.4}$, $N = 22$, $r = 0.72$	...	0.8 to 2.8
Densely rimed columns	$V = 1.1L^{0.56}$, $N = 13$, $r = 0.79$	$V = 1.8M^{0.11}$, $N = 13$, $r = 0.49$	$M = 0.033L^{2.3}$, $N = 13$, $r = 0.78$	0.02 to 0.27	0.8 to 2.0
Densely rimed dendrites†	$V = 0.62D^{0.33}$, $N = 10$, $r = 0.54$	$V = 1.2M^{0.16}$, $N = 9$, $r = 0.68$	$M = 0.015D^{2.3}$, $N = 9$, $r = 0.90$	...	1.8 to 4.0
Densely rimed radiating assemblages of dendrites*	$V = 1.1D^{0.12}$, $N = 14$, $r = 0.23$	$V = 1.3M^{0.06}$, $N = 13$, $r = 0.34$	$M = 0.039D^{2.1}$, $N = 13$, $r = 0.92$	...	0.8 to 2.8
Unrimed side planes	$V = 0.81D^{0.99}$, $N = 10$, $r = 0.77$	...	...	...	0.4 to 1.2
Aggregates of unrimed radiating assemblages of dendrites or dendrites*	$V = 0.8D^{0.16}$, $N = 28$, $r = 0.20$	$V = 1.1M^{0.08}$, $N = 27$, $r = 0.15$	$M = 0.073D^{1.4}$, $N = 27$, $r = 0.91$	...	2.0 to 10.0
Aggregates of densely rimed radiating assemblages of dendrites or dendrites	$V = 0.79D^{0.27}$, $N = 27$, $r = 0.55$	$V = 1.3M^{0.15}$, $N = 25$, $r = 0.69$	$M = 0.037D^{1.9}$, $N = 25$, $r = 0.88$	...	2.0 to 12.0
Aggregates of unrimed radiating assemblages of plates, side planes, bullets, and columns§	$V = 0.69D^{0.41}$, $N = 31$, $r = 0.91$	$V = 1.2M^{0.07}$, $N = 19$, $r = 0.36$	$M = 0.037D^{1.9}$, $N = 19$, $r = 0.84$	...	For V vs. D, 0.2 to 3.0; for V vs. M and M vs. D, 1.0 to 3.0
Aggregates of unrimed side planes†	$V = 0.82D^{0.12}$, $N = 23$, $r = 0.29$	$V = 1.2M^{0.14}$, $N = 21$, $r = 0.63$	$M = 0.04D^{1.4}$, $N = 21$, $r = 0.78$	...	0.5 to 4.0

See text for definitions of maximum dimensions in ambiguous cases. Fall speed V is given in meters per second; mass M, in milligrams; and maximum dimension D or L, in millimeters; N is the number of datum points, and r is the correlation coefficient for the relationship.

*The probability that the correlation between V and D or V and M will be accidental is greater than 0.1.

†The probability that the correlation between V and D will be accidental is greater than 0.1.

§The probability that the correlation between V and M will be accidental is greater than 0.1.

To produce Figure 4.9 we selected data for those crystal types observed on Mt. Seymour. Crystals with broad branches (P1c) also occurred frequently on Mt. Seymour (45% of the storms, Table 3.5). Davis (1974) derived empirical formulae for a variety of crystal shapes and sizes including P1c. Over a range of ice-crystal diameters of 0.01 to 1.0 mm the appropriate relationship was:

$$\text{Terminal velocity } (\text{cm}\cdot\text{s}^{-1}) = 1.39 \times 10^{2}\ d^{\,0\ 748} \qquad (4.1)$$

where d is diameter in cm. Thus terminal velocity of such crystals 1 mm in diameter would be 24.8 $\text{cm}\cdot\text{s}^{-1}$. That agrees closely with Kajikawa's data of Fig. 4.7b which suggest a terminal velocity of about 30 $\text{cm}\cdot\text{s}^{-1}$.

Needles were the most common crystal shape on Mt. Seymour (Table 3.5), but we can find no estimate of their terminal velocity in the literature. If the empirical formula derived by Davis (1974) for solid columns with a length/diameter ratio > 2.0 is applied (assuming a length of 1000 μm, Fig. 3.23a), we obtain a terminal velocity of 2.4 $\text{m}\cdot\text{s}^{-1}$. That value is slightly higher than those documented by Locatelli and Hobbs (1974) for rimed columns (see Fig. 4.8).

Kuz'min (1963) used photographic techniques to determine that the average angle of snowflake fall was 4° from the horizontal. In the Sierra Nevadas, Court (1957) found that only 2% of the precipitation fell during calm periods. Because snowfall in coastal areas is dominated by rapid

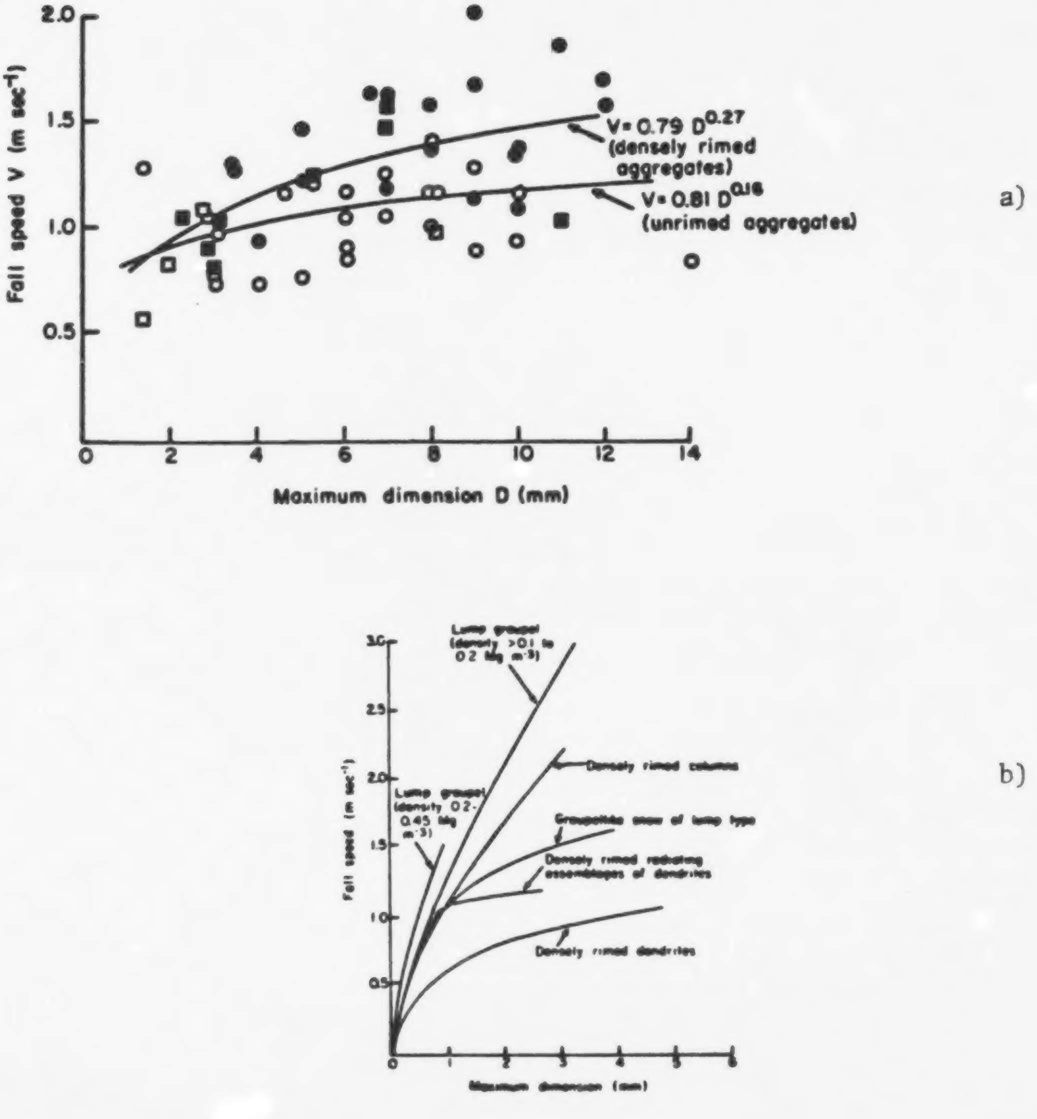

Figure 4.9 a) Fall velocities as a function of maximum dimensions for unrimed aggregates of radiating assemblages of dendrites (open circles), densely rimed aggregates of radiating assemblages of dendrites (solid circles), unrimed aggregates of dendrites (open squares), and densely rimed aggregates of dendrites (solid squares) (from Locatelli and Hobbs 1974: 2192).
b) Best fit curves for fall velocity versus maximum dimensions of single solid precipitation particles of various types (modified from Locatelli and Hobbs 1974: 2191).

orographic lifting (Ch. 3.2.1), snowfall there will also be accompanied frequently by winds. However, it seems probable that snowfall on the coast is more nearly vertical than the value reported by Kuz'min (1963), because relatively high terminal velocities are probable given the crystal and aggregate types present (Table 3.5, Fig. 4.8). Computations of terminal velocities must make restrictive assumptions. We are aware of no generalized relationship between wind speed, ice crystal size and density, and angle of attack. We can conclude only that the angle of snowfall in coastal forests will show greater departure from horizontal than 4°.

There are at least three practical effects of the diagonal pattern of snowfall. First, wind may blow snow far from the point of condensation, delivering snow to areas that might otherwise not receive that amount of snow. The fact that snow input does not stay put is treated in Chs. 7 and 8. Second, the effective interception surfaces of trees become angular rather than simple vertical or horizontal projections. Relationships of wind to the interception capabilities of trees is discussed in Ch. 7.3-7.6. The effect of wind on the kind and form of vegetation present is not discussed. Third, wind modifies the simple relationship of Figure 4.1 to produce more complex relationships of the type illustrated in Figure 4.2.

4.3 Summary

A more comprehensive summary of the influence of abiotic features on snow delivery is presented in Chapter 5. The statement here simply reviews major points noted in Chapter 4.

All three major effects of wind on snow delivery (orographic, crystal modification, and imposition of diagonal vectors) are likely important in British Columbia. Snowfall in coastal regions derives largely from frontal systems experiencing orographic lifting which implies that winds accompany many snow storms. Effects of wind on crystal structure relate primarily to the character of deposited snow and imply that new snow in coastal forests should be relatively dense. We note two implications to the angular vectors contributed by wind. First, given the strong orographic effect on precipitation in coastal British Columbia, more snow may be delivered to sloping ground than to level areas (Fig. 4.2). That would invalidate the simple model proposed for 'slope-effect' (Fig. 4.1); the effect likely will be complicated in coastal areas. Second, the relatively high terminal velocities of snow and ice aggregates in coastal mountains, suggests that wind effects on distribution during the delivery phase will be less than in drier regions. Nevertheless, it is apparent that changes in wind speed or turbulence can profoundly affect snow delivery. Komarov (1963) concluded that the deposition of blowing snow was dependent on the cubic or greater power of the difference between initial and final velocities. That observation suggests that small reductions in wind speeds can produce large changes in snow deposition. The influence of forest cover on wind speed (Ch. 4.2, Fig. 4.3) will serve to increase the deposition of snow in forested areas.

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5. ABIOTIC EFFECTS ON SNOW DELIVERY - SUMMARY

Snow interception in forests can be described simply as "snow meets trees". The purpose of Chapter 5 is to summarize briefly the first half of the statement; that is, to describe the production of snow in south coastal British Columbia and how that snow meets trees. It reviews and summarizes the relationships treated in detail in Chapters 3 and 4 by considering three broad topics:

1) under what conditions does snow form in south coastal British Columbia (Ch. 5.1),

2) what kinds of snow will fall (Ch. 5.2), and

3) how will that snow be delivered to surfaces on the ground (Ch. 5.3).

By summarizing abiotic conditions generating snow and its probable delivery patterns, Chapter 5 establishes broad boundary conditions to the complex phenomenon of interception. We can then offer some predictions regarding interception. These predictions are presented verbally in Chapter 6 and are quantitatively evaluated for individual tree and stand attributes in Chapters 7 and 8. The present chapter is meant to provide a succinct summary of those abiotic features previously treated that are most relevant to interception. Comments on the predictability of specific relations are included. For brevity, the summary statements are presented in point form; the appropriate sections of Chapters 3 and 4 are cited.

5.1 Conditions Producing Snow

South coastal British Columbia is mountainous and the conditions producing snow must consider phenomena at three scales (Ch. 3, Introduction): i) large-scale, synoptic factors determining the characteristics of air masses involved; ii) the dynamics of air motion over mountains; and iii) the microphysics of clouds and precipitation formation.

1. At the largest scale, coastal British Columbia is the preferred track for cyclonic disturbances, many of which originate with, or are influenced by, the Aleutian Low. During winter, maximum moisture transport across the Pacific Ocean occurs between latitudes of 50-55°N (Ch. 3.1). As a result, Vancouver Island and the adjacent mainland receive larger amounts of precipitable water in winter than any other area of Canada (Fig. 3.5).

2. The global patterns are corroborated by local measurements. On Mt. Seymour during the winter, frontal systems represented 83% of all storms and 84% of those

producing snow (Table 3.1). Most systems involved moist, maritime air masses, only 8% of the fronts involved drier, Arctic air masses.

3. Whether a storm produces snow is to a great extent dependent on the temperature and moisture content of the air masses involved. In south coastal British Columbia, the probability of snow at lower elevations is very unpredictable for two reasons: i) the diversity of air masses potentially involved (Fig. 3.4); and ii) the diverse trajectories these masses can follow when approaching the coast (Fig. 3.6). Different mixtures of air masses and their trajectories exhibit different probabilities of occurring and of producing snow. First approximations to these probabilities are summarized in Table 3.2.

4. Some snowfall at elevations above about 500 m is virtually assured in all winters. That is true because most air masses are moisture-laden when they arrive at the coast (Figs. 3.5 and 3.6). The empirical lapse rate of about $7^{\circ}C \cdot km^{-1}$ ensures sufficient adiabatic cooling for snow to fall (Fig. 3.1).

5. In coastal British Columbia vertical air motion and adiabatic cooling produced by frontal storms are substantially augmented by orographic lifting.

Conditions favouring orographic lifting include: i) strong winds perpendicular to mountain ridges; ii) air masses moist in depth and containing existing cloud; iii) lapse rates near neutral stability; and iv) significant vertical relief. Each of these conditions is met in south coastal British Columbia.

6. The favourable conditions for orographic lifting produce a strong orographic component to precipitation. Snowfall in a given storm increases with elevation (Fig. 3.8) and the orographic effect increases with increasing storm size (Eq. 3.1).

7. We thus expect snowfalls of different characters to occur at low and high elevations. Under those rare conditions when the air mass is sufficiently cool for snow to fall near sea level, frontal conditions will often prevail. Snow particles frequently will fall through relatively calm air, following a large angle of approach from horizontal, and producing soft, often deep snow. At higher elevations, orographic lifting and considerable air motion will produce a lower angle of approach by smaller snow aggregates. The broad associations of elevation with snow particle size and angle of approach have opposing effects on the probability of interception, and also influence the utility of different crown measurements predicting interception at different

elevations (Ch. 6).

8. Beyond the statements of point 5 through 7, little of an unequivocal nature can be stated about events occurring at the mesoscale of air motion over mountain ridges. Given that the air masses involved are usually moist, and the orographic effect effect is pronounced, snowfall should be predictable using only wind speed and air temperature (Eq. 4.4). The probability of snowfall is broadly predictable (Fig. 3.10), but the elevation at which snow falls is predictable only if the mean freezing level is known (Fig. 3.11).

9. Although the adiabatic lapse was nearly constant on Mt. Seymour (Eqs. 3.2 and 3.3), regression equations of the proportion of precipitation which fell as snow versus elevation exhibited a large standard error. The lack of precision occurred because of the variable nature of air masses involved, which produced different freezing levels. First approximations of freezing levels by storm type are presented in Table 3.4; they were often non-normally distributed even within broad storm types. One result is that different relations with elevation were observed in different years (Fig. 3.9). That in turn makes it difficult to provide reliable guidelines regarding favourable elevations for ungulate winter ranges.

10. Processes occurring at the third scale (cloud microphysics) also reduce the reliability of predictions regarding the occurrence and magnitude of snowfall. Neither moisture (Fig. 3.5) nor temperature (Fig. 3.1) are likely to be limiting to the formation of ice crystals. Although the concentrations of cloud condensation nuclei are lower in maritime air (Fig. 3.14), they are activated relatively quickly (Ch. 3.3.1) and any potential limitation is likely to be of short duration. However, concentrations of CCN are influenced by changes in air mass, wind speed, and direction so that short-lived limitations to snowfall are possible (Fig. 3.15).

11. The concentrations of CCN that act as ice-forming nuclei (IN) do not increase monotonically with decreasing temperature in coastal areas (Fig. 3.18). Because the effects of supersaturation (over ice) on IN concentration are independent of temperature (Fig. 3.17b), the variability in IN concentration with respect to temperature appears to result from variable moisture contents in the air masses. Thus, even though the lapse rate is nearly constant, the ability to predict snowfall as a function of elevation is poor: the formation of ice-forming nuclei is controlled by moisture content as well as by temperature.

12. Of the four factors potentially limiting to ice-crystal formation (cloud condensation nuclei, ice-forming nuclei, moisture, and temperature), temperature is most often limiting at lower elevations of coastal British Columbia (Ch. 3.3.1). Temperature and moisture naturally interact in the microphysics of a rising air mass. The variability in air masses produces variability in elevation effects (Table 3.4), but the moist nature of most air masses and strong orographic effect will almost always allow both supersaturation (Fig. 3.1) and ice crystal formation (Fig. 3.17) at higher elevations (above 500 to 1000 m).

13. Growth of precipitate is rapid in the air masses common to south coastal British Columbia. Maritime clouds contain droplets of greater diameter than do continental clouds (Fig. 3.19a) and growth rates by coalescence are more rapid (Fig. 3.19b). During winter, initial growth of precipitate by vapour diffusion is more common (Ch. 3.3.2). Under temperature and moisture conditions prevailing in south coastal British Columbia, these rates are also rapid. Once nucleation has occurred, only a few minutes are required for crystals to attain sufficient mass to begin falling (Figs. 3.20 and 3.24). Time for snow crystal growth is rarely limiting. Thus, despite rapid orographic lifting, the precipitate forms and falls on the windward sides of the mountains. As with other

potentially controlling processes, the microphysics of ice crystal growth suggest that snowfall will always occur at higher elevations, but be poorly predictable at lower elevations.

5.2 Kinds of Snow Produced

The abiotic conditions under which snow forms, determines the kinds of snow produced by governing: crystal type or shape, dimension, mass, density, and the likelihood of aggregate formation and riming. These snow particle attributes, in turn, influence interception processes by their influence on terminal velocities, angle of approach, adhesion, and cohesion. They also influence the density of snow, thus weight of the snow load and the likelihood of overload throughfall and wind erosion (Ch. 6).

14. Snow crystal shape is governed by the temperature and moisture conditions during crystal growth (Figs. 3.21 and 3.22). Those relationships indicate that elementary needles and needle aggregates should be the most common crystal type in coastal areas, and they are (Table 3.5). Sector crystals (P1c) were the second most common crystal type, which is also expected given the relationships governing crystal growth. Dendritic crystals, which form at temperatures of about -15°C, were less common. These latter crystal types could occur frequently under those

colder conditions when snow is produced at lower elevations.

15. Maximum dimensions and mass of individual, unrimed crystals are relatively small, with mean maximal dimensions of 1.0 to 2.5 mm and masses of about 0.1 mg or less (Figs. 3.23 and 4.8).

16. Because conditions in south coastal British Columbia favour crystal aggregation, dimensions of individual crystals is of lesser significance than in colder, drier areas. More than half the mass of solid precipitation reaching the ground in the Cascade Mountains was in the form of aggregates. Ice particles are unlikely to be frequently limiting to crystal aggregation in coastal British Columbia (Fig. 3.26), thus, temperature is the major factor determining aggregation (Ch. 3.3.2). Warmer temperatures (below freezing) encourage larger aggregates (Fig. 3.25), and we expect frequent, large aggregates in coastal mountains of British Columbia providing appropriate crystal types are present.

17. All expectations considering the frequency and size of aggregates are met. Aggregates occur with frequencies exceeding 60%. Because needle shapes predominate, most aggregations remain small (3-4 mm in diameter, Fig. 3.27). Dendritic types occur in about 30% of the

storms (Table 3.5) and produce radiating assemblages of dendrites with mean dimensions of 6 to 9 mm (Fig. 3.27), but up to 10 to 12 mm (Fig. 4.8). Mass is related to dimension by power laws, and dendritic aggregates would attain masses of about 2 mg (Fig. 3.28, Table 4.1).

18. Conditions in south coastal British Columbia also encourage riming which increases crystal and aggregate mass. The rapid rates of lifting of moist air ensure the presence of liquid water. Theoretical calculations from data of Fig. 3.24 and Table 3.5 indicate that growth by vapour diffusion of only 1 minute would produce dimensions of common crystal types large enough to permit the onset of riming (Figs. 3.29 and 3.30). Many crystals collected on Mt. Seymour were rimed and graupel occurred in 62% of the storms (Table 3.5). Riming and graupel formation in coastal British Columbia would often follow the wet growth regime with slow freezing rates, thus producing lower concentrations of air bubbles. That in turn should produce dense snow and ice particles (0.2 - 0.45 $g \cdot cm^{-3}$) with high terminal velocities (Table 4.1, Fig. 4.8), and relatively dense, new-fallen snow.

5.3 Delivery of Snow

Delivery of snow to surfaces on the ground is a function of the kind of snow produced (Ch. 5.2), wind speed, and the angle or slope of the intercepting surfaces. All three of these interact. Speed of vertical air motion is a function of slope. The abiotic conditions producing particular kinds of snow are also covariate with wind speed. Forests on the ground will modify wind speed.

19. Two broad classes of wind speed were noted (point 6): i) calm conditions in frontal weather, generally at lower elevations; and ii) windy conditions at higher elevations where orographic lifting would modify otherwise calm conditions. Especially at higher elevations, much of coastal British Columbia's snowfall will occur at wind velocities exceeding 1 $m \cdot s^{-1}$. Factors contributing to windy conditions (other than conditions favouring orographic lifting, point 5) include: i) the general pattern of geostrophic winds (Ch. 4.2.1); ii) local winds created by channeling in rugged topography; iii) greater vertical velocity caused by steep slopes (Ch. 4.2.2); and iv) the convergence of markedly different air masses (Ch. 3.1). Maritime occluded fronts would produce turbulent mixing and exhibited the highest probability of producing snow of all storm types on Mt. Seymour (Table 3.2).

20. The presence of wind has several important effects. Some of these have been noted: i) facilitation of riming (point 18); ii) production of rapid rates of crystal

growth through ventilation (Fig. 3.24a); and iii) precipitation on windward sides of mountains (point 13). There are three further broad effects: i) contribution to orographic effects; ii) modification of ice crystals; and iii) reduction of the snow particles' angle of approach. Geostrophic and other winds contribute to orographic effects by breaking up low level inversions and encouraging a near-neutral lapse rate (Ch. 4.2.2), and because precipitation is directly proportional to vertical velocity (Eq. 4.4). Whereas cold air is mostly deflected away by mountain ridges, warm air is deflected up and over ridges increasing in velocity as it goes. One result is that higher rates of precipitation are encouraged at higher elevations.

21. Moderate wind speeds encourage crystal and aggregate growth (Ch. 3.3.2), but higher speeds mutilate crystal aggregates and break them into their elemental, needle-shaped components. Elementary needles and broken branches occurred frequently in storms on Mt. Seymour (Table 3.5). Such modification increases density of newly-fallen snow, and modifies adhesive and cohesive forces, thus wind erosion or overload throughfall of intercepted snow (Ch. 6).

22. A major effect of wind on interception is the introduction of a diagonal vector or low angle of

approach to snowfall. Because of their high drag and low density, most snow particles have terminal velocities only 10 to 15% of those of raindrops of similar dimension (Ch. 4.2.2). Thus, small increases in wind speed can lower the angle of approach considerably. Given the high probabilities of aggregation (point 16) and riming (point 18) in coastal British Columbia, snow particles there have greater mean dimensions, mass, and densities than in many other areas. As a consequence they also have relatively high terminal velocities of 0.3 to 2.0 $m \cdot s^{-1}$ (Table 4.1). Their angle of approach will generally exceed the value of 4° from horizontal documented for colder, drier snow, but will still be sufficiently low that vertical components of interceptor surfaces will be important.

23. A corollary of the observation that small increases in wind speed lower the angle of approach is that small reductions in wind speed significantly increase the likelihood of deposition of the snow particles from the air stream. The wind velocity profile is a function of stability of the air (as measured by the vertical temperature gradient) and surface roughness (Eq. 4.2). Because air in coastal British Columbia appears relatively stable (Eqs. 3.2 and 3.3), surface roughness exerts the dominant control. Forest cover generates relatively high values of surface roughness (Fig. 4.3)

and requires only very low wind speeds (< 0.01 $m \cdot s^{-1}$) to generate turbulent flow (Fig. 4.5). The interaction of forest cover with wind speed will have pronounced effects on interception processes. At low to moderate wind speeds the turbulence generated by forest cover will encourage deposition of snow from the air stream, at higher wind speeds the effects will be more complicated and mediated by the shape, slope, and flexibility of intercepting surfaces (Chs. 6 and 7). Very simply, we expect higher wind speeds to favour gravity over adhesion or cohesion, thereby reducing interception.

24. Effects of slope on rates of snow delivery interact with both wind speed and rates of snow production (the latter two are themselves covariate). The simplest model suggests that, in calm air, for geometric reasons alone, less snow should fall on slopes (Fig. 4.1). As a result, adhesive and cohesive forces will be less effective and snow on slopes is more likely to be redistributed and less likely to accumulate. As increasing wind speed lowers the angle of approach and increases rates of snow production, the predictions change (Fig. 4.2). The combined interaction is complex, particularly for small surfaces such as those comprising tree crowns. Slope is important as it provides a vertical component to the interceptor surface. The importance of vertical area depends on the angle of approach of the snow particles (a

function of wind speed). Moderate slopes and wind speeds should increase the efficiency of interception. Greater wind speeds and slopes favour gravity over adhesion and cohesion, particularly on smaller surfaces which can be shaken by wind. Specific predictions for smaller surfaces are prescribed in Chapter 6.

III. EFFECTS OF FORESTS ON SNOW DEPOSITION

6. CONCEPTUAL FRAMEWORK AND DEFINITIONS

The study of snow interception by forests has a long history in North America. Church (1912) noted the practical importance of the observation that less snow accumulates under the forest canopy than in the open. Thirty years of collecting semi-quantitative data followed his pioneering work. Efforts were descriptive and largely free of theory. In the late 1940's, Kittredge (1948, 1953) developed ideas concerning snow redistribution in forest openings and the fate of intercepted snow. Many of his concepts have dominated thinking until the present day. Apparent conservatism and concentration on description left major principles largely unquestioned (thus unchallenged), despite early, thoughtful attempts by Miller (1962, 1964) and others to relate observations back to underlying processes.

Research which has questioned principles has exposed omissions and inconsistencies in long-accepted relationships. The present theoretical uncertainty makes the task of the reviewer and synthesizer an awkward one. Interpretation of results is also complicated by the recurring problem of confounding factors. Most research on snow is performed by practical hydrologists who are concerned primarily with consequences of run-off from maximal, spring snowpacks. Consequently, less attention has been given to distinguishing among the individual processes of interception, redistribution, and melt - all of which contribute to snowpack development. Short term practicality has aggregated processes into black boxes.

The present review begins with definitions and a conceptual framework that attempts to differentiate processes involved in snow interception. That is the purpose of this chapter. Individual processes are examined using available data in Chapter 7. The approach continues by aggregating processes affecting interception by single crowns into stand measurements, until a highly integrative, non-causal description is attained (Ch. 8). That description is equivalent to the black box normally measured. It will be evident that aggregation is equivalent to obfuscation. Our hope is that by first reviewing concepts and processes, understanding of individual processes will be increased.

6.1 Definitions

Crown measurements.--The use of crown measurements in forestry practice has a long history. Count Reventlow of Denmark utilized ratios of tree height to crown width in the early part of the 19th century (Van Slyke 1964). Today, foresters routinely measure and evaluate crown characteristics yet, after 150 years, there is no standardized terminology nor widely accepted methodology for crown measurement.

Caution is necessary when evaluating and comparing published crown measurements in any but the most superficial manner. Crown measurements vary widely depending on the worker's definitions and methods. That observation is particularly true with respect to the most widely used measurement, "canopy cover" or "crown closure". Canopy cover often refers solely to the proportion of the ground overlain by tree canopy. However, some workers also incorporate the degree to which an individual tree's crown is 'complete'. Furthermore, it is well documented that values of canopy cover measurements are highly dependent upon the means of measurement employed. Discrepancies between canopy covers determined by various means have been reported by Dodd et.al. (1972), Rochelle (1975), and Majawa (1977).

The terminology and definitions used in this report follow:

1) Crown Closure = Canopy Cover - the proportion of the ground surface encompassed by vertical projections of the outer edges of tree crowns. This measurement is better suited to stands and is usually used in that context. For analysis we have had to use an analogue for individual trees, it is termed projected horizontal area (Ch. 7.6).

2) Crown Density - commonly the mean biomass per m^3 of the crown. We have used an approximation of mean length of primary branches per m^3 of crown.

3) Crown Completeness - the proportion of the sky obliterated by tree crowns within a defined angle (or determined with a described instrument) from a single point. This is a point measurement obtained with such instruments as a moosehorn (Robinson 1947), spherical densiometer (Lemmon 1956), or camera. It combines reductions in cover resulting from both the absence of tree crowns and from holes within tree crowns.

4) Mean Crown Completeness - a stand measure determined from a number of crown completeness measures.

Snowpack and snow deposition.--In the literature the term "interception" is often used uncritically. We believe it is correctly used only when it refers to that amount of snow or proportion of a snowfall which does not reach the ground during a given storm. It can be approximated by the difference between new snow in the open and new snow under the canopy. Too often interception is used to refer to the difference between snowpack in the open and snowpack under the canopy. Differences in snowpack arise from a host of factors including true interception, the fate of intercepted snow, melt rates, and redistribution of snow by wind. Interception has no clear meaning when applied to the snowpack, but becomes increasingly meaningful the more recently the snow has been deposited.

Throughfall and interception.--In this review *interception* refers to snow which remains in the tree canopy after cessation of the storm. That proportion of the snow which is not intercepted is considered *throughfall*; throughfall may occur through the crown of individual trees or in openings between individual trees. Conceptually, canopy throughfall can be further subdivided into *true throughfall*, that which falls unimpeded through holes in the canopy; *ricochet throughfall*, that which ricochets off canopy parts and bounces or sifts to the ground; and *overload throughfall*, that snow which is initially intercepted but which falls from the crown before the storm is over.

Because true throughfall of snow occurs when snow crystals fall unimpeded through the tree's canopy, the amount of true throughfall depends solely upon the 'effective' size and number of holes in the crown. Larger snowflakes are more likely to contact the sides of a crown opening of any particular size. More important is the orientation of the canopy opening relative to the angle of the snowfall. Canopies with holes aligned parallel to the angle of snowfall will intercept the least snow (Fig. 6.1a). Holes are reduced in size as snow accumulates. Effective size refers to openings as the snowflake responds to them, not necessarily as current canopy measurements record them. That statement is particularly true for crown closure, less so for crown completeness. An important task of the researcher is to develop a canopy measurement which reflects effective hole

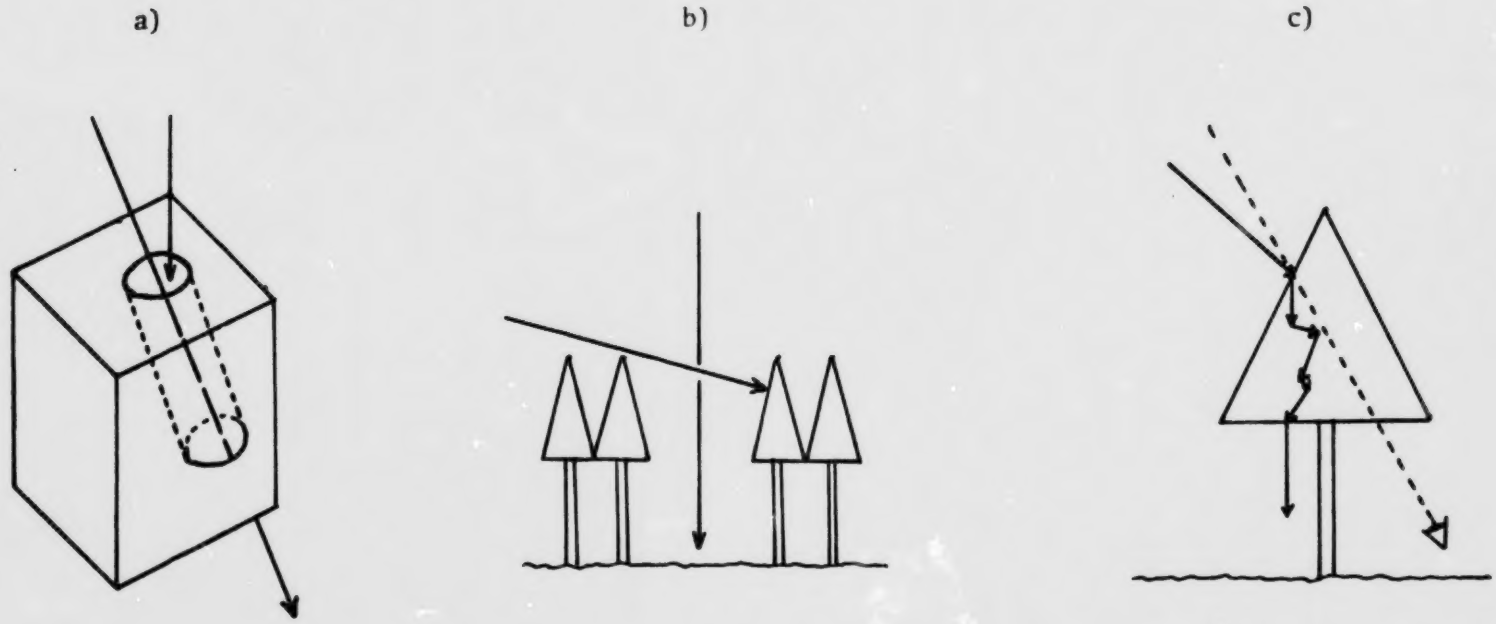

Figure 6.1 a) Schematic representation of interception and true throughfall within a tree crown.
b) Schematic representation of interception and true throughfall within a stand.
c) Schematic representation of richochet throughfall. Note that canopy measured by line of sight does not reflect the snow crystal's erratic path.

size.

The same process also occurs on a larger scale. Forest openings or clearings are effectively largest to vertically falling snow. Smaller angles of approach by snow crystals decrease effective opening size (Fig. 6.1b). One implication is that wetter, heavier snowfall common in coastal regions should experience greater true throughfall than lighter, drier snow because openings are effectively larger in the former case. Within a single crown, however, true throughfall may well be less for coastal snows because of larger aggregate size or windy conditions which reduce the angle of approach.

Some snow which enters the canopy strikes needles, twigs, and branches but bounces off them repeatedly before dropping to the ground. Crown or canopy completeness probably does not adequately reflect ricochet throughfall because the erratic track taken by the snow particle is not measured by line of sight canopy measurements (Fig. 6.1c). More important are the density and depth of the canopy (which determine the potential number of hits), and the tendency for the particle to stick which is dependent upon the strength of cohesive and adhesive forces. Because cohesion and adhesion are both strongest at warmer temperatures (Ch. 7.1) we expect ricochet throughfall to be important only at cold temperatures.

Tree canopies have a physical limit to the amount of snow that can accumulate on or in them. This maximum is termed maximal snow load and is measured in kg or kg per unit area of SWE (snow water equivalent). Once the maximum is attained, further snowfall drops from the crown and is operationally defined as overload throughfall if it occurs during the storm and mass transport of intercepted snow if it occurs after the storm.

The maximal snow load is reached when cohesive and adhesive forces are exceeded by forces of wind or gravity. Maximal snow load and overload throughfall are therefore dependent on those features influencing cohesion and adhesion (effective roughness, wind, temperature), the size and orientation of intercepting surfaces present, and the flexibility of the intercepting surfaces which allows them to bend under snow load and thereby direct the gravitational vector more effectively relative to the intercepting surface.

We thus expect relationships governing maximal snow load and overload throughfall to be very complicated and dependent upon tree species, individual tree morphology, and storm type.

6.2 A Conceptual Framework

The conceptual framework used in subsequent chapters owes much to the work of Miller (1964, 1966). He provided detailed 'word models' of interception and attempted to quantify his descriptions with published information. Our approach to the word model differs superficially from Miller's in its general structure; it interprets some concepts differently and distinguishes three broad stages during interception. Our efforts to quantify particular relationships have benefited by having more published information available than did Miller (Chs. 7 and 8). Nonetheless, Miller's approach is somewhat more comprehensive as he included more information from areas having drier snow, and utilized data on rainfall as well as snow.

The three broad stages we consider are: i) delivery of snow to the canopy; ii) deposition of snow from the air stream; and iii) transport of snow from the canopy during a storm. Some of the following statements are based on our interpretation of processes, not on data. The word model presented here is meant to provide an overview; specific portions are elaborated in subsequent chapters.

6.2.1 Delivery of Snow to the Canopy

The rate of delivery of snow to a canopy depends upon the intensity of precipitation, and the structure of the canopy relative to the trajectory of the snow crystals (Figs. 6.1a and 6.2). Factors influencing intensity of precipitation have been reviewed (Chs. 3 and 4). Snow crystal trajectories differ above a canopy from those that occur in the open. The review of Kuz'min (1963) indicated that snow particles in the open fall with an average angle from the horizontal of 4°. When snow particles enter the zone of slower wind speed just above the tree crowns, their angle of approach becomes more vertical, but their paths are altered by eddies produced by the rough upper crown surface and forest openings. Delivery

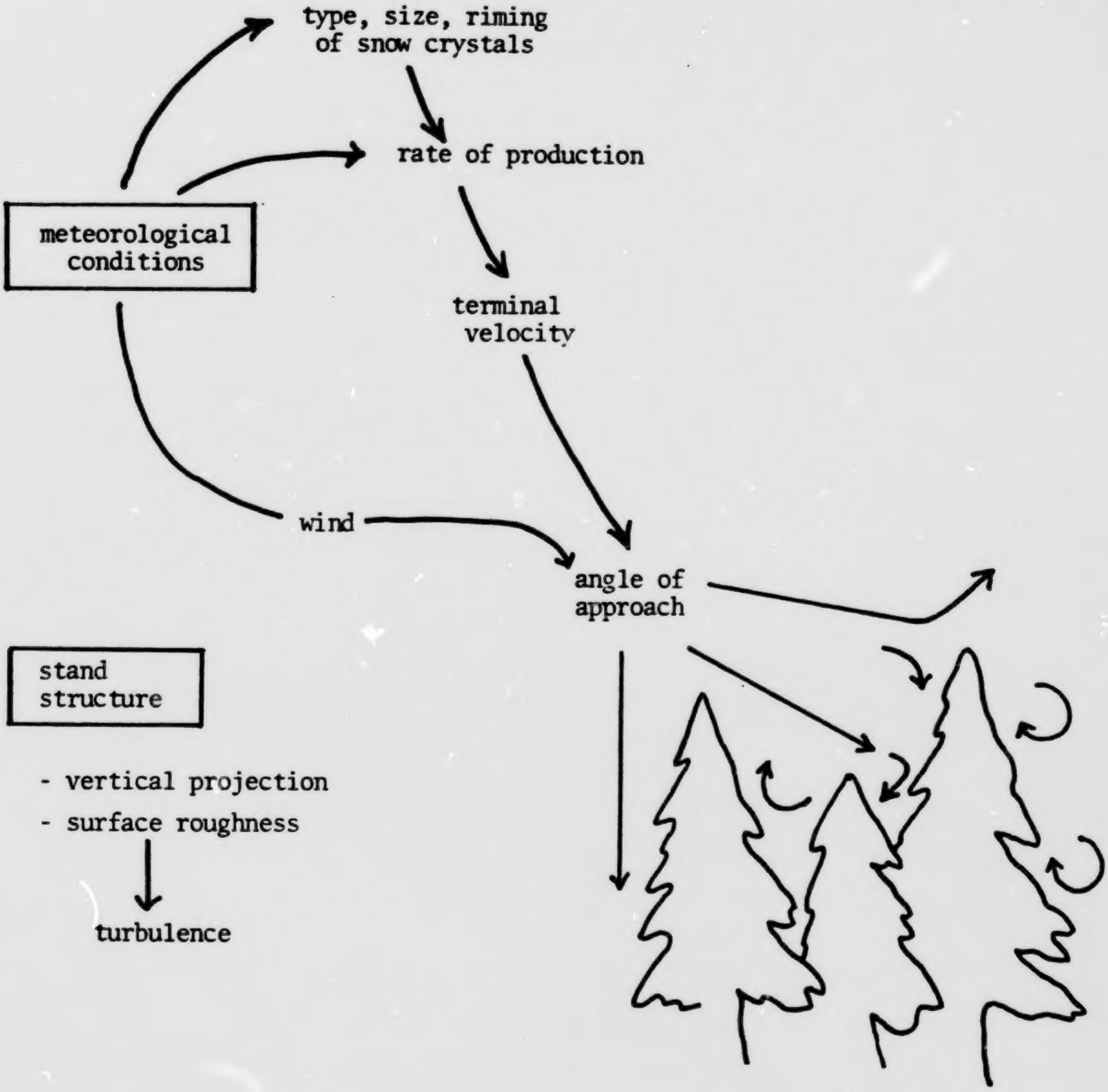
type, size, riming of snow crystals
rate of production
meteorological conditions
terminal velocity
wind
angle of approach
stand structure
- vertical projection
- surface roughness
turbulence

Figure 6.2 Schematic representation of factors affecting the delivery of snow to a forest canopy.

of snow to a forest canopy has been measured a few times by precipitation cans at crown level (e.g., Davis 1939), but results have not been analyzed to evaluate the role of turbulence or roughness of the forest surface.

The low terminal velocities of most snow particles (Ch. 4.2.2), coupled with the turbulence produced by the canopy surface, encourages a low angle of approach and has the effect of increasing the potential intercepting surface of an incompletely closed stand (Fig. 6.3). Furthermore, the turbulence above a rough surface tends to draw precipitation particles down into the crown (documented for fog droplets by Hori 1953). As a result, crown attributes considered only as a horizontal plane are usually poor predictors of snow interception (e.g., Ch. 7.6). At no time, except in dead calm, is throughfall governed merely by the openings in a two-dimensional vertical plane.

There are several implications of this description of snow delivery to forest canopies:

1) The heterogeneous, vertical projection of a forest canopy, together with its surface roughness (and accompanying turbulence), will make it an effective interceptor. Forestry practices that alter the canopy surface (e.g., selective logging of dominants) will alter the efficiency of interception; rougher canopies will intercept more snow.

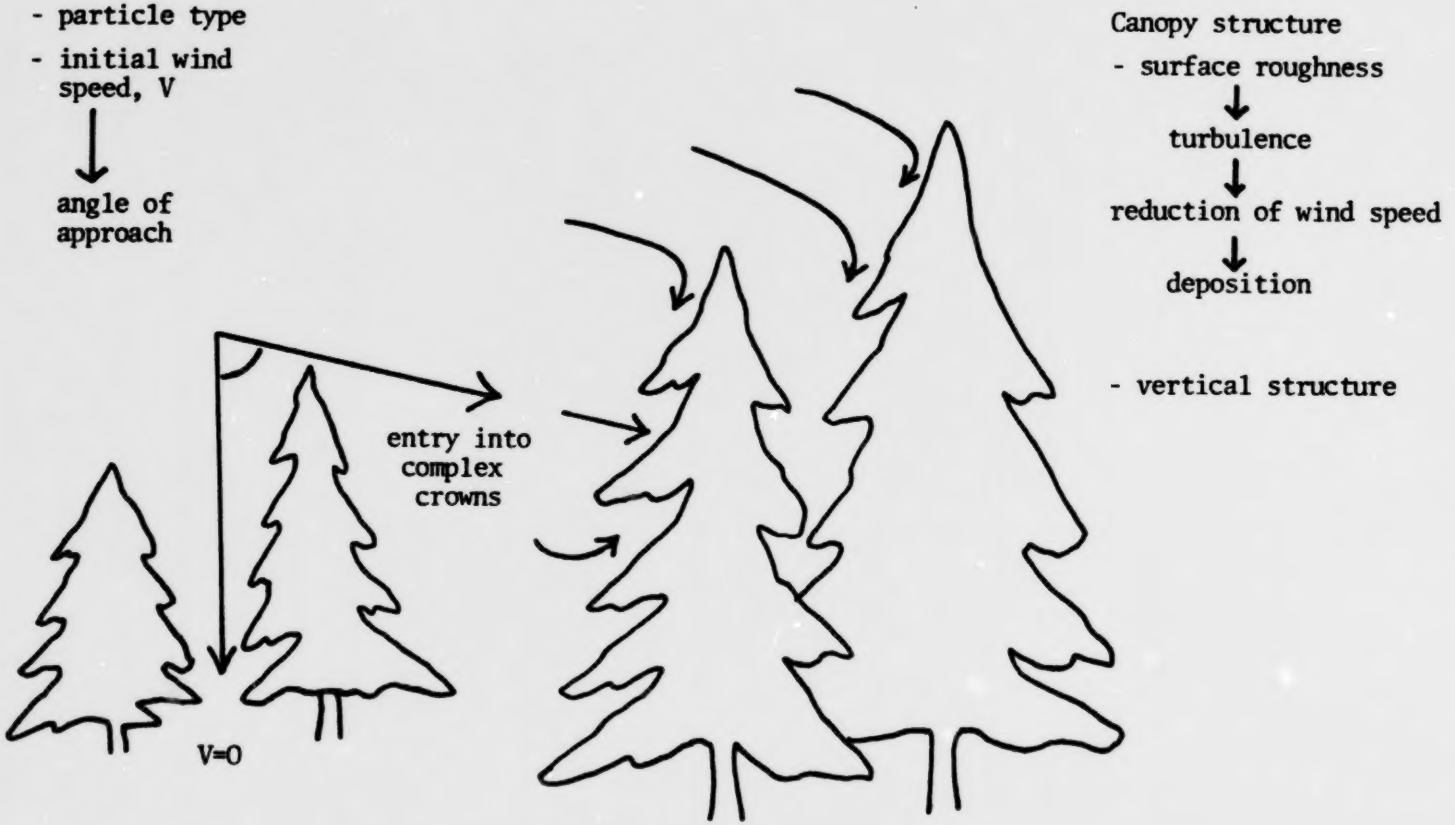
- particle type
- initial wind speed, V
angle of approach
V=0
entry into complex crowns
Canopy structure
- surface roughness
turbulence
reduction of wind speed
deposition
- vertical structure

Figure 6.3 Schematic representation of factors affecting the deposition of snow into a forest canopy.

2) Measurements of crown attributes related to interception should include a vertical component. The desirability of a vertical component increases with increasing wind speed. Rowe and Hendrix (1951) reported that, in wet storms of low wind speed, the annual snow catch at crown level approximated that in cans placed in small openings. Frontal weather systems are common in coastal British Columbia (Table 3.1), and often have calm air below the frontal inversion. During such weather, snow falling through quiet air near the earth's surface approaches the ground vertically. Thus, at low elevations snowfall may be nearly vertical. The vertical vector is strengthened by the density and size of snow crystals and their aggregates in warm, maritime areas (Ch. 3.3.2), and their relatively high terminal velocities (Ch. 4.2.2). At higher elevations orographic lifting (Ch. 3.2.1) becomes more important and introduces a larger diagonal vector to snowfall. Thus the utility of measurements of crown characteristics may change with elevation. This distinction between high and low elevations is broad; there can be strong directional wind shears below as well as above frontal inversions.

3) Wind speed has an obvious effect on delivery of snow to intercepting surfaces, primarily through its effect on the angle of approach. Low wind speeds should increase interception above that in still air, higher wind speeds

will tend to reduce interception in a complex fashion depending on the angle of the intercepting surface relative to the wind (Chs. 4.1 and 7.6).

4) Turbulent eddies occur at larger scales than in the canopy itself (e.g., topography, Fig. 4.6; forest-openings, Ch. 8.3) and also affect delivery of snow particles at those scales. Miller (1964) noted that turbulent eddies introduced by topography may account for the extreme snow loads sometimes reported for trees on lee slopes.

6.2.2 Deposition of Snow from the Air Stream

For snow to be intercepted by trees it must first be dropped out of, or deposited from, the air stream. The initial crystals must adhere to tree surfaces, after which cohesive forces dominate in the accumulating snow load. Abiotic conditions and characteristics of interceptor surfaces influence each stage.

When the flow of snow and air encounters the porous forest canopy the velocity is decreased. Because deposition depends on the difference between the cubic or greater power of initial and final wind speeds (Komarov 1963), small decreases in wind speed should result in a large deposition of snow (Ch. 4). Wind speed itself, or features that reduce wind speed (e.g., surface roughness), will have a major influence on the amount of snow initially deposited from the air stream (Fig. 6.3). In still air, particularly with dry snow, some snow particles will penetrate the canopy without alighting on branches or foliage (true throughfall). Under such conditions, crown cover as measured in a two-dimensional horizontal plane could reflect accurately the interceptive potential. Because completely still air is rare, most snowfall has a low angle of approach and the vertical component of the crown becomes important to interception. Thus moderate increases in wind speed of 1-2 $m \cdot s^{-1}$ should increase interception by complex crowns (Fig. 6.3). The lower angle of approach would allow snow to penetrate the inner areas of the crown or encounter vertically-oriented surfaces more directly. As snow load increases when snow is deposited on snow, branch flexibility will determine the angle of the receptor surface relative to the snow crystals' angle of approach.

Considering only deposition of snow, we have five predictions:

1) Interception of snow on simple, sloping surfaces will decrease with increasing wind speed.

2) Interception of snow by complex surfaces (e.g., tree crowns) will increase with moderate increases in wind speed.

3) At low wind speeds, and over a wide range of slope, interception will increase with increasing slope (providing angle of delivery is still diagonal).

4) Particular combinations of initial branch angle and branch flexibility may increase the effective intercepting surface as snow loads increase.

5) The best predictors of interception will incorporate a vertical component of the tree crown.

These predictions are evaluated in Chapter 7, all appear correct.

Once deposited from the air stream, snow must adhere to the crown for interception to occur. The major factors influencing adhesion are temperature and characteristics of the intercepting surfaces such as texture, angle, shape, and flexibility (Fig. 6.4). Wind can increase adhesion by impacting snow particles onto rough surfaces.

Adhesion need not occur; falling particles could simply bounce off surfaces (ricochet throughfall) and likely do under dry conditions. If snow particles do not adhere easily to foliage on contact, they penetrate deeply into the stand. Miller (1955: Fig. 7) reported that about twice as much snow as solar radiation penetrates a forest of the same stem density. Because adhesion is dependent on a film of liquid water, it is strongly temperature-dependent. The optimal

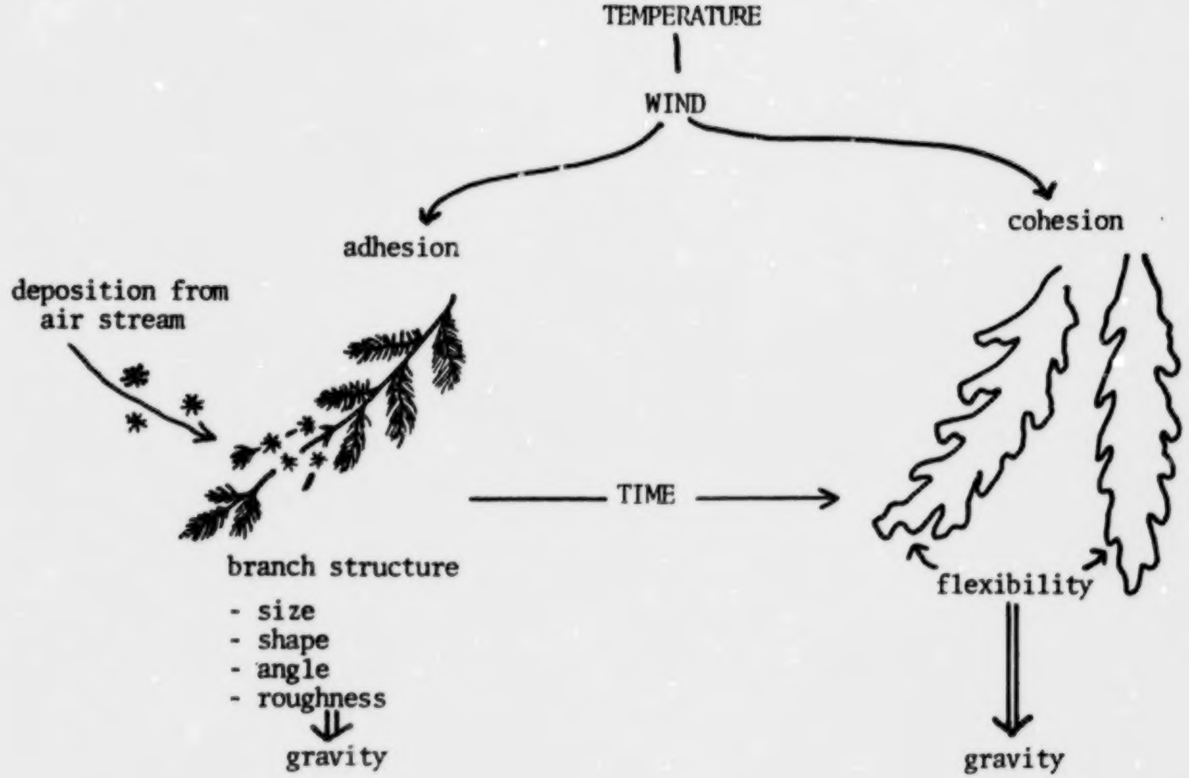
TEMPERATURE
WIND
adhesion
cohesion
deposition from air stream
TIME
branch structure
- size
- shape
- angle
- roughness
gravity
flexibility
gravity

Figure 6.4 Schematic representation of factors affecting the accumulation of snow loads during snow storms.

conditions for adhesion of snow should occur just below freezing. The major influence of surface texture may be its influence on liquid water retention. At this stage of interception the surface characteristics of angle, shape, and flexibility still influence the relative surface area exposed to a particular angle of delivery. They also influence the outcome of the competing forces of gravity and adhesion. Shape is important because gravity will compete more effectively with adhesion on rounded than on flat surfaces. Similarly, the steeper the angle the less effective is adhesion in counteracting gravity. Surface flexibility will determine the angle under snow load; the greater the angle the less effectively snow will be retained against gravity (Fig. 6.4). In nature the potentially independent effects are combined. Baldwin (1957) reported that, when heavily loaded, the branches of fir, spruce, and hemlock droop and let the snow slide off, while the stiffer branches of pine continue to hold it. In large storms, the flexibility of pine needles may be countered by the stiffness of pine branches (see also Ch. 7.6.5).

Unless snow is wet and sticky, the net effect of greater wind speeds on adhesion appears to be negative (at least in terms of the resulting snow load). Although Minsk (1961) reported a greater frequency of adhesion of snow to mesh panels, lesser amounts of snow accumulated at lower wind speeds. In large part, the result occurs because vibration by wind favours gravity over adhesion or simply dislodges the snow. We therefore expect that one influence of increasing wind speed will be to reduce snow loads more effectively on steeper slopes or more flexible surfaces (in this instance the effect influences cohesion as well).

Once snow has adhered to the canopy, cohesive forces between snow particles determine the effective interception (providing gravity or wind do not disrupt adhesion to the intercepting surfaces). The major factors influencing cohesion are the same as those influencing adhesion. At temperatures prevailing in coastal British Columbia, the major process encouraging cohesion is sintering through the translocation of vapour or liquid water to refreeze as solid bonds. Interlocking of dendritic structures is more common at lower temperatures (Figs. 3.21 and 3.22). We, therefore, expect cohesive forces also to be most effective just below freezing. The interaction of surface characteristics such as angle and flexibility with wind and gravity, are similar to those for adhesion. Specifically, increasing angle and flexibility will decrease snow loads, expecially at higher wind speeds. Surface shape changes the relative likelihood of failure of adhesive and cohesive forces. Adhesion must counteract gravity only perpendicular to a flat, level plane, and cohesion is then the dominant force in accumulating snow loads. On rounded surfaces, gravity acts parallel to the extreme edges and adhesion is more likely to fail. Wind has an additional positive effect on cohesion because it can encourage vapour translocation thus sintering.

Several predictions can be extracted from the word model:

1) Snow loads will decrease as temperatures decline below freezing, and adhesive and cohesive forces are lessened.

2) Ricochet throughfall will be uncommon in coastal British Columbia (effective adhesion), but at colder temperatures (drier conditions) crown depth will still be important to interception (not all particles adhere on first impact).

3) Under conditions of warm, moist snow, small surfaces will accumulate snow efficiently on a unit-area basis (effective cohesion and adhesion).

4) Flat surfaces will accumulate snow more effectively than round surfaces (adhesion more effective).

5) Snow load will decrease with increasing angle of the surface, especially at higher wind speeds. The decline should be abrupt where gravity overcomes adhesion or cohesion.

6) Low wind speeds at temperatures near freezing will encourage accumulation of snow loads by favouring vapour translocation and cohesion.

7) Depending on initial angle of orientation (Ch. 6.2.1)

more flexible surfaces will bear lighter snow loads (corollary of 5).

8) The most effective crown measurements of interceptive potential should incorporate some measurement of branch density, size, and flexibility.

These predictions also are evaluated in Chapter 7. The first seven were found to be true. Insufficient data were available to separate unequivocally the potential influences of different crown attributes on interception.

6.2.3 Transport of Snow from the Canopy

Snow caught by a canopy during a storm may not remain in the canopy until the end of the storm (Fig. 6.5). For some questions it does not matter where the once-intercepted snow went, for others it is critical. There are at least five routes by which snow can leave a canopy during a storm. To a large extent they are inseparable: i) vapour transport from snow; ii) vapour transport from melt water; iii) stem flow and dripping of melt water; iv) wind erosion; and v) sliding from branches without the intervention of wind.

Intercepted snow has no source of radiative or convective heat sufficiently strong to raise its surface temperature and vapour pressure much higher than the temperature and vapour pressure of the air. During a snow storm the air is cold and

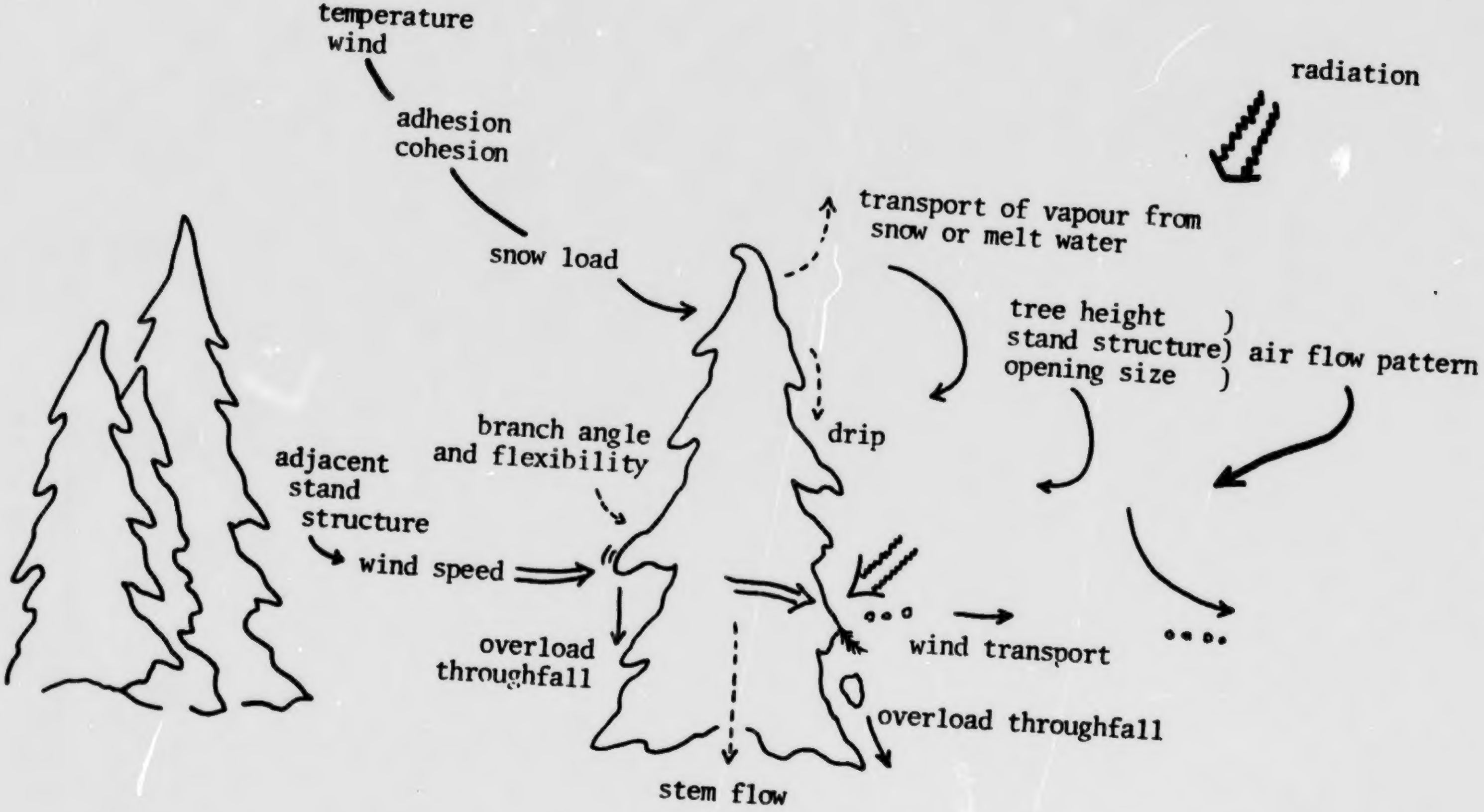
temperature
wind
adhesion
cohesion
snow load
radiation
transport of vapour from
snow or melt water
tree height)
stand structure) air flow pattern
opening size)
branch angle
and flexibility
drip
adjacent
stand
structure
wind speed
overload
throughfall
wind transport
overload throughfall
stem flow

Figure 6.5 Schematic representation of factors affecting the removal of intercepted snow from tree canopies during a snow storm.

humid and cannot itself provide a large supply of sensible heat for evaporation of intercepted snow. As a result, vapour transport from intercepted snow during a storm is likely to be small enough to be well within the measurement error of most studies. By imaginative treatment of data not directly relevant to the question, Miller (1966) estimated daily rates of evaporation during a storm to be 0.7 to 1.0 mm SWE. Actual rates are likely smaller because he assumed the air was dry enough to permit evaporation. For similar reasons, primarily insufficient heat, vapour transport from melt water during a storm is also likely to be very small (less than 1 cm of snowfall).

Transport as liquid water will exceed losses in vapour form. Stem flow and dripping of melt water are encouraged by humid, cloudy weather just above freezing. Such conditions prevail in coastal snow storms and favour melting over evaporation. Delfs (1958) and Eidmann (1959) reported that stem flow on Picea abies in Germany attained measurable quantities; Eidmann estimated it as somewhat less than 1% of precipitation. Values are likely higher in more maritime climates where large snowfall, greater heat supply, and higher humidity favour melting. Rowe and Hendrix (1951) found stem flow in ponderosa pine to be about 3% of total snowfall, but values ranged from 1 to 6%. In snow storms over six winters, stem flow averaged 1.5 mm per storm; to produce that much melt water required 12 ly heat per storm.

The losses of intercepted snow through melted and evaporated water are unlikely to exceed 2 mm SWE during a storm, and are therefore primarily of interest to those individuals studying specific processes. There is one important exception to that statement. Many studies of interception make comparisons between snow in the open and snow under the canopy. They must therefore make assumptions about the degree to which the open area is the repository of snow initially intercepted by trees. The preceding observations suggest that during a storm, most snow removed from trees will be removed by wind erosion or sliding from branches.

During snow storms, wind affects intercepted snow by shaking supporting surfaces, gradually eroding masses of snow, or blowing larger fragments off their supports and transporting them down wind (Fig. 6.5). Hoover (1960) asked the vexing question "How much of the snow in openings has blown off the foliage of surrounding trees"? If snow initially caught on foliage is subsequently blown into openings sheltered from wind, forest openings become a sink for snow falling over a larger area. Interception is overestimated.

The processes governing wind erosion of intercepted snow are similar to those governing its interception. Provided wind speeds are moderate, temperatures near freezing will discourage erosion by favouring adhesion and cohesion. However, snow that develops strong cohesion can accumulate in masses large enough to be blown off by wind; Morey (1942) observed that heavy accumulations are more likely to be blown off than light ones. Cramer (1960) made similar observations when evaluating effects of helicopter 'blasts' on snow loads. Data reviewed in Chapter 7.3 suggest that the critical wind speed for isolated trees in a relatively warm storm is about 3 $m \cdot s^{-1}$ (10.8 $km \cdot h^{-1}$). Winds at that speed blew off about half the intercepted snow (Fig. 7.7). In stands where a dense, continuous canopy separates the trunk space from free air, wind velocities are lower and less wind transport of snow occurs. Sakharov (1949) estimated that 25 to 30% more snow reached the ground under a single-storied spruce stand subject to wind action than under a two-storied stand in which the understory experienced less wind.

Terminal velocities and surface to volume ratios of clumps of blowing snow determine how far they will be carried in the air stream. Turbulence of the wind in the trunk space determines how much snow can be carried as suspended load. These, together with the height from which intercepted snow is blown, will determine whether snow reaches an adjacent opening or the ground immediately below the canopy. Higher crowns provide a longer period of travel while falling and permit greater velocities. The observation that eddies smaller than 100 m are the prevailing size in a snow storm (Rogers and Tripp 1964) suggests a critical dimension for openings receiving excess snow (see also Ch. 8).

Given the relatively warm, moist air in coastal winters both adhesion and cohesion will be effective. However, the critical wind speed of 3 $m \cdot s^{-1}$ also was estimated under warm, moist conditions (data JGFES 1952). Except during calm, frontal weather, some snow intercepted in coastal regions will be transported by wind. Because cohesion will normally be effective, that snow is more likely to be dislodged in clumps, and clumps large enough that large, downwind distances are unlikely. The removal of smaller fragments will also occur. Because these will be transported under humid conditions, they are unlikely to lose much mass to evaporation and may be carried significant distances (subject to turbulence). These observations, plus the relatively high density of newly-fallen, coastal snow, suggest that although wind will transport intercepted snow, most of the transport will be over short distances. Transport will be most effective from trees with steeply sloping, flexible surfaces.

Intercepted snow may also slide off tree branches without the intervention of wind if the branches provide a sloping or unstable surface. Temperature again influences the effectiveness of adhesion and cohesion in opposing gravity, but any disturbance that releases the potential energy of a loaded branch can be important. The disturbance may be the impact of snow clumps falling from above or the rapid melting of bonds holding snow to a sloping surface. Because relatively little heat is required to cause sufficient melting for large clumps to fall off, the process could occur commonly in coastal British Columbia. Whereas wind transport of snow to openings will produce overestimates of interception (by artifically increasing snow in the open), overload throughfall of snow to the ground beneath the canopy will underestimate interception. Unless these processes are measured, or trees are continuously weighed, the estimate of interception is in doubt.

The word model suggests several predictions:

1) In coastal conditions, transport of intercepted snow as liquid water or vapour during a storm could be as high as 7% of total precipitation, but is more likely to be 1 to 2%.

2) Measured snow loads will decrease with increasing wind speed.

3) Measured snow loads will decrease with small increases in radiation or temperature.

4) There should be an optimum size of forest openings for catching and retaining wind-transported snow. The size will be a function of adjacent tree height, because that will influence the pattern of air flow.

5) Wind transport of intercepted snow will be less on the coast than in drier regions.

6) Accurate estimates of interception comparing open and

forested areas will be thwarted by wind transport and overload throughfall or snow sliding from branches.

Predictions 2) and 3) are evaluated and found true in Chapter 7. Predictions 4) and 5) are evaluated and found to be qualitatively correct in Chapter 8. Prediction 6) is indirectly evaluated in Chapter 7 and found to be true. Prediction 1) could not be evaluated.

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7. FACTORS AFFECTING INTERCEPTION IN SINGLE TREES

7.1 Physical Forces

Other than gravity there are only two physical processes which determine whether snow striking the canopy will be intercepted or continue to fall to ground as ricochet or overload throughfall. These are adhesion, the process of snow sticking to the tree, and cohesion, the process of snow sticking to snow already present. It is difficult to separate or measure the two processes independently, but is is helpful to separate them conceptually.

Adhesion.--For snow interception to occur, snow must initially rest on or adhere to the tree and leaf surfaces. The adhesive process is greatly facilitated by the presence of a thin, water film on the tree. Adhesion is therefore greatest at warmer temperatures. Vegetation warmer than freezing melts the first snow to hit it thereby producing free water. Later snowfall cools the water freezing the snow to the tree. The opposite situation can lead to the same result (a thin water film); when the tree is colder than freezing and the snow is warm and wet. Wet snow "clinging" to trees is a common observation and one that is frequently cited in the literature (see review in Miller 1964).

If both the tree and the snow are well below freezing, adhesion does not benefit from the stickiness of a water film.

Cold snow particles must then lodge mechanically in the canopy. The degree of roughness of leaf, twig, branch, and bole surfaces are of particular importance. Form of the crown and flexibility of the foliage and branches influence the area available to support adhering snow. For example, time lapse photography of Lull and Rushmore (1961) showed that white pine needles accumulated snow only at the base of the fascicles before the needles were bent into a platform.

Wind can significantly affect adhesion by forceably driving snow particles into the tree thereby increasing adhesiveness through impaction. Minsk (1961) reported higher wind speeds increased adhesion of snow to a mesh panel.

Cohesion.--After the first layer of snow adheres to the canopy, further accumulation occurs by cohesion among the snow particles. Processes of cohesion are similar to those producing snow crystal aggregates (Ch. 3.3.2). They include the interlocking of dendritic points and the sintering of snow particles whereby water is translocated and refrozen to produce solid bonds between particles. Clearly, the former is of more significance at temperatures of -12 to -17°C and the latter in warmer conditions (Fig. 3.21). Rime mixed with snow can increase cohesion through cementing of snow load.

There are very few data directly relating temperature and wind speed to the major physical processes involved in interception (adhesion and cohesion). Minsk (1961) and Kuriowa (1962) provide some data. We know of only one detailed experimental study analyzing the effects of temperature and wind on interception (The Japanese Government Forest Experiment Station [JGFES], 1952). The Japanese researchers used physical models as well as individual trees, reduced the problem to separate, discrete questions, and controlled for a large number of otherwise confounding variables. Much of the following discussion is based on their work.

7.2 Temperature

To simulate the differences between wet and dry snow, researchers dropped wet and dry sawdust on Cryptomeria japonica (cedar) branches. The maximal load of wet sawdust was approximately double that of dry sawdust. Figure 7.1 illustrates the role of wetness independent of temperature in strengthening adhesive and cohesive forces.

Both simulated and real cedar (Cryptomeria) branches caught or intercepted more wet sawdust, and generally retained more than 50% of the sawdust dropped onto their surfaces. The greater efficiency of interception by the model is a result of its lower flexibility and shallower slope (discussed more fully in Ch. 7.6). We consider wetness as analogous to warmer temperatures. Recall that temperature effects on interception efficiency act through the production and freezing of free water (Ch. 7.1).

Figure 7.2 illustrates the effect of air temperature on

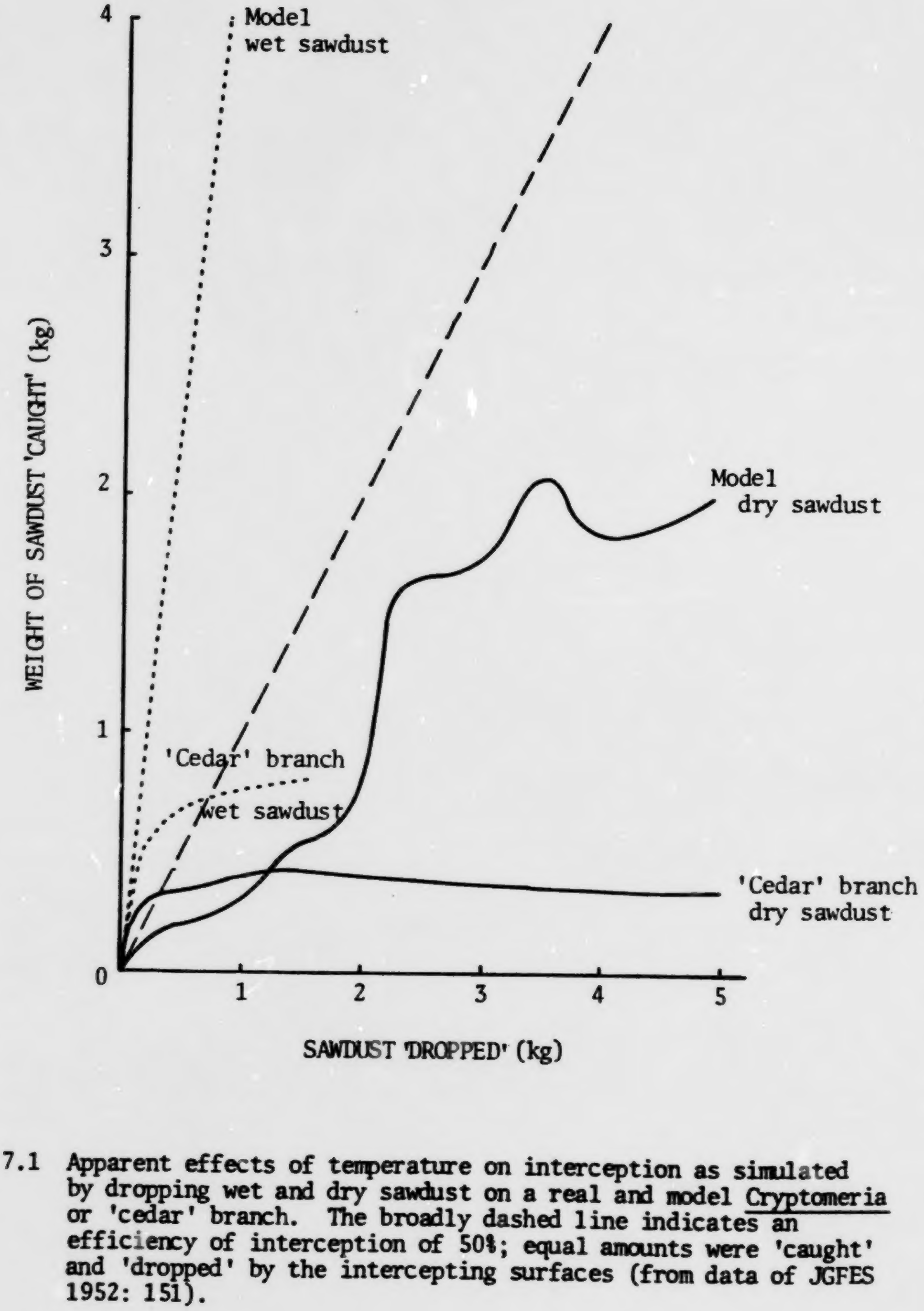


Figure 7.1 Apparent effects of temperature on interception as simulated by dropping wet and dry sawdust on a real and model _Cryptomeria_ or 'cedar' branch. The broadly dashed line indicates an efficiency of interception of 50%; equal amounts were 'caught' and 'dropped' by the intercepting surfaces (from data of JGFES 1952: 151).

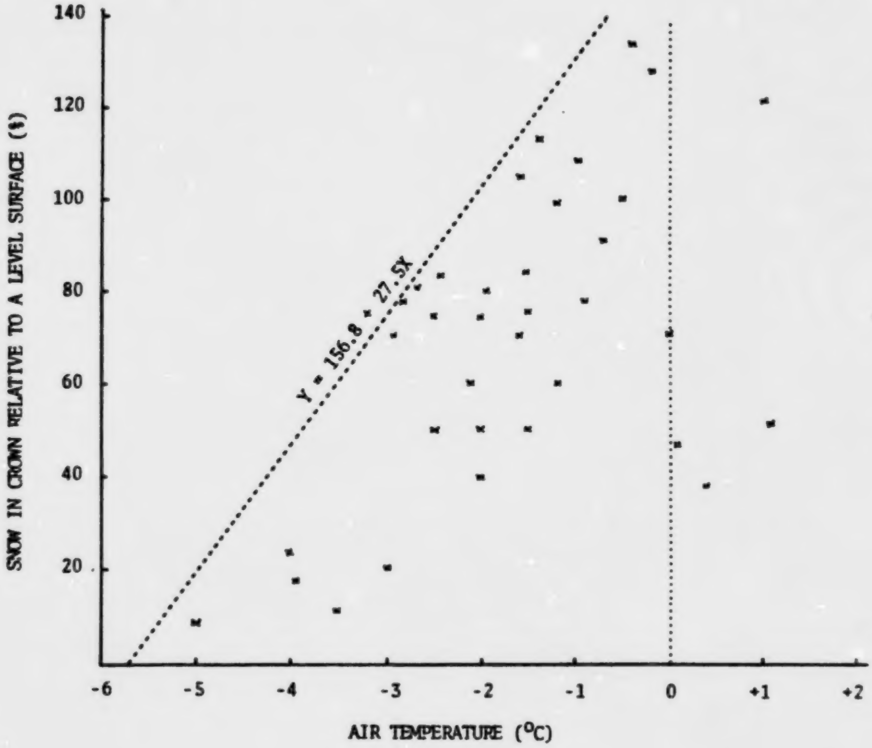


Figure 7.2 Per cent of snow in a *Cryptomeria* crown relative to that on a level surface as a function of air temperature during snowfall (from data of JGFES 1952: 143). The regression line represents the envelope which includes all data.

the rate at which a single Cryptomeria tree increased in snow load relative to snowfall on a level surface. Weights of snow on the tree (as a percentage of snowfall in the open), were always small at low temperatures (about -5°C), moderate to heavy at temperatures just below freezing, and declined markedly at air temperatures above freezing. At temperatures just below freezing, the weight of snow in the crown exceeded that of the horizontal projected area of the crown and relative snow load exceeded 100% (Fig. 7.2). That occurred because snowfall is diagonal and the entire length of the crown can intercept snow when adhesion and cohesion are effective.

JGFES derived an upper envelope for the relationship by regressing snow load as a percentage of snowfall (Y) on air temperature below freezing (X) including all observations (Fig. 7.2). Most observations followed a trend approximated by $Y = 130 + 30X$ (where X is negative). Near freezing both adhesive and cohesive forces are increased. The observed relationship suggests that together these forces increase the efficiency of interception by approximately 30% of snowfall for each degree Celsius of warming less than freezing. The five storms in which intercepted snow amounted to less than 20% of that on the ground were colder (temperatures of -3°C or lower), and presumably the snow was drier (Fig. 7.2).

Cohesion generally increases most rapidly at temperatures approaching freezing (Ch. 3.3.2) and the underlying relationship of Figure 7.2 is likely curvilinear. We do not have the raw data and cannot test the form of the relationship. As temperature increases towards freezing, the two major metamorphic processes encouraging cohesion have opposing actions. The mechanical linking of irregular flakes is reduced; sintering or bonding by vapour translocation is increased. The data of Figure 7.2 suggest that latter process is dominant in encouraging cohesion.

Because the Japanese workers obtained continuous weights of a snow-laden tree during individual storms, case histories of interception were obtained. They illustrated three such histories (JGFES 1952: 143, Fig. 6). When the air temperature remained between -3° and -4°C for the entire storm, the snow load closely followed the curve of accumulated storm precipitation. In another storm, in which the air temperature was initially just above freezing but fell to -1°C during most of the storm, snow load accumulated rapidly and uniformly. Relative to snow load on the ground it outweighed snow loads of other storms and attained 22 kg in 10 h. The rate of increase in load, 1 mm water equivalent per hour, was a large fraction of the rate of snowfall. The observations again document the importance of temperature to interception.

JGFES (1952) also evaluated the influence of temperature on adhesion and cohesion by measuring the amount of snow held on boards 1, 2, 4, and 8 cm wide relative to that held on a 16 cm board. The relationship for horizontal boards of varying widths illustrates that the effect of width on maximal snow load increases with decreasing temperature (Fig. 7.3). At a

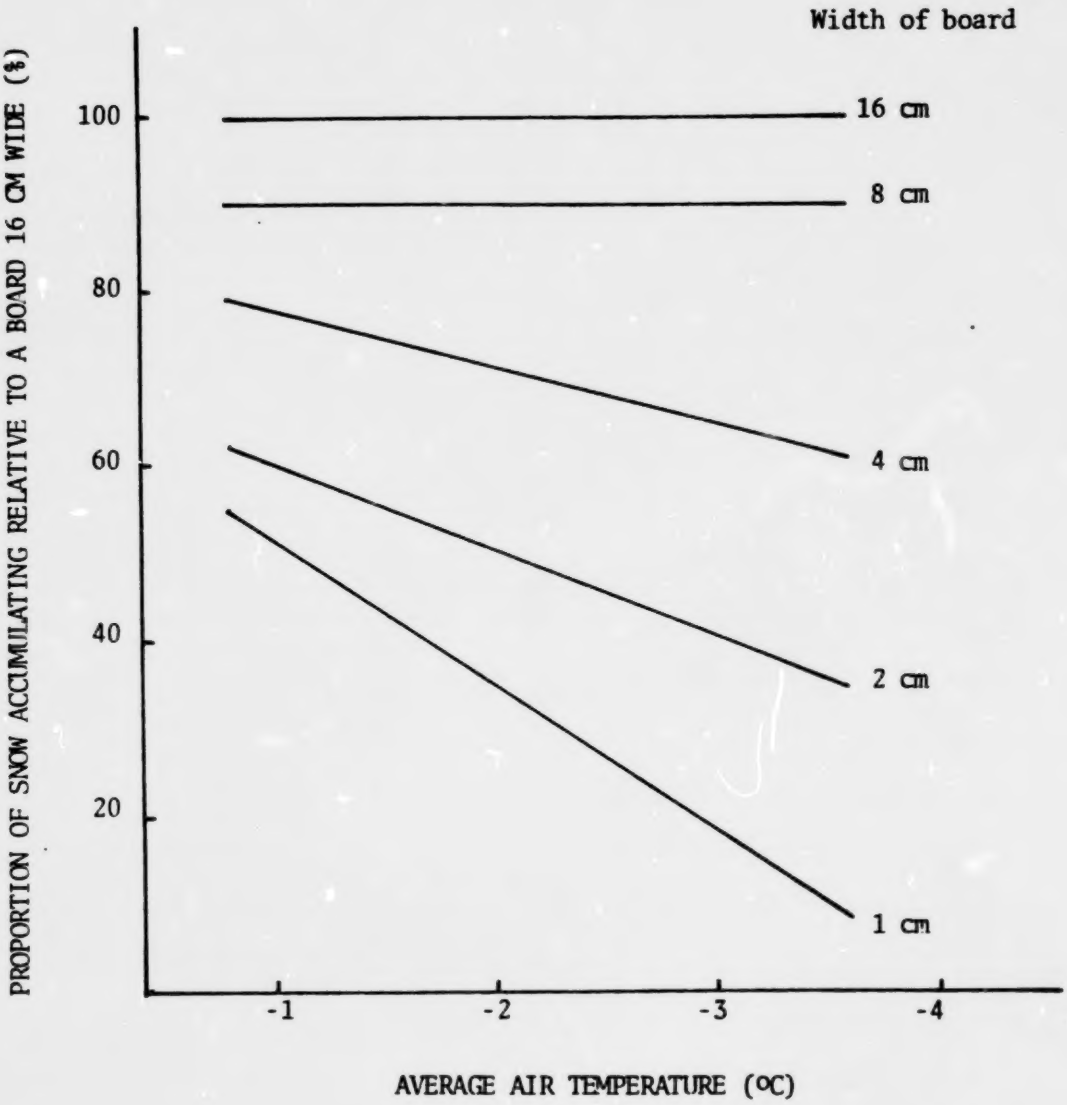


Figure 7.3 Effect of air temperature during snowfall on the accumulation of snow per unit area on horizontal boards of varying width (from data of JGFES 1952: 119).

temperature of -3°C, a board 1 cm wide held only about 20% of the snow per unit horizontal area held by a board 16 cm wide; at -1°C, the 1 cm wide board held 55% of the maximal snow load per unit area held by 16 cm board. Temperature has a relatively greater effect on the snow load of small surfaces than on large surfaces due to the greater role adhesion and cohesion play in holding snow onto small surfaces.

Contrary to Miller (1964) we ascribe much of the increased snow load at warmer temperatures to increased cohesion. Adhesion has been shown in laboratory studies (Kuriowa 1962) and in field experiments (Minsk 1961) to be temperature-dependent, probably because of the importance of thin, liquid water films. But once the snow has adhered cohesive processes take over, and cohesion also is enhanced at temperatures close to freezing (Fig. 7.2). Similarly, in the case histories of interception reported above, the greatest initial rates of interception occurred when air temperature was above freezing at the onset of the storm. The 1-cm board of Figure 7.3 could attain 55% of the load per unit area of a 16-cm board only if cohesion permitted the snow load to build up.

Most additional observations on the potential influence of temperature on interception are anecdotal. Baldwin's (1957) comment that high-density snow is more effectively intercepted than low-density snow presumably refers to warm, wet snow in the former case. Heikenheimo (1920) noted that vegetation initially warmer than freezing encouraged adhesion, whereas Tikhomirov (1938) regarded the reversed differential (trees cold and snow warm and wet) as the most important factor favouring heavy deposits of crown snow. Both cases would involve a film of liquid water. Similarly, Bennett (1959) observed that sometimes snow falling after glaze adds to the snow load.

Where snow falls from warm maritime air (as in Japan and much of western North America) it often falls through a layer of air that has a temperature above freezing. The Japanese experiments (JGFES 1952, Watanabe and Ozeki 1964) occurred in air temperatures up to +2 and +3°C, respectively. Miller (1955: 25) reported that in the crest region of the Californian Sierra Nevada snowfall occurs at air temperatures up to 35°F (+1.7°C) and a third of it falls at temperatures above freezing. Provided the snowfall is intense, some individual snowflakes will survive the warm temperatures to arrive under conditions favourable for adhesion and cohesion. For a time, heavy snowfalls will accumulate faster than they melt, but small increases in air temperature will start melting of snow masses, especially at their points of attachment. The result is more often mass transport of snow as it slides off branches than complete melting of the intercepted snow load. Temporal patterns of accumulated snow loads documented by JGFES (1952) indicate large decreases after supply of relatively small amounts of heat. The pattern suggests sliding of partly-melted snow from branches.

Sliding, partly-melted snow requires relatively little heat, considering how much snow is transported, primarily because mass transport is encouraged by the potential energy stored when snow lies on sloping branches or imposes elastic loading on them. In one storm (JGFES 1952: Fig. 11) during a day with light wind and air temperatures rising from -2° to 1°C, the snow load on a Cryptomeria tree decreased linearly from 25 mm SWE at 1000 hours to 2 mm at 1500 hours. Averaging several continuous storms that continued from night into daylight hours, indicated that the mean load of 8 mm SWE that persisted from 0400 to 0800 hours decreased after sunrise at an hourly rate of about 20% of the snow load still on the tree (see Fig. 7.9). This high rate of transport within a snowfall (1 to 2 mm SWE h^{-1}) probably reflects both increasing insolation and warmer air (see also discussion of Fig. 7.10, Ch. 7.4).

Implications to coastal British Columbia are: i) that adhesive and cohesive forces will frequently be effective because of the warm, moist weather, thereby producing large, crown snow loads; ii) other factors being equal, these forces will be most evident in trees composed of small surfaces; and iii) accumulated snow loads will often be shed rapidly, particularly on trees with pendant branches (Ch. 7.6.2 and 7.6.3).

The latter point has serious implications for efforts calculating interception as the difference between snow in the open and under the canopy. Time of sampling (relative to when snow is shed) is clearly important, but shedding may also produce shallower but denser snow under the canopy relative to the same SWE of uncompacted snow in the open. If apparent interception cannot be measured before shedding begins, measurements should consider potential differences in densities in the open and under the canopy. Densities under coniferous canopies are commonly greater than in the open, even for newly fallen snow (Bunnell and Jones 1985) and differences can be dramatic in the upper 48 cm (Bunnell et al 1985). Over the entire snowpack, however, densities may be higher in the open than under canopies (Bunnell et al. 1985).

7.3 Wind

Wind affects apparent interception in three major ways. It drives the snow onto the canopy with force, thereby increasing adhesive and cohesive forces. This effect presumably could be counteracted by the effect of increasing wind speed reducing the contact time between snow particles impacting intercepting surfaces at an angle and the surfaces themselves. Effective time for cohesion and adhesion could be reduced, particularly at colder temperatures. Secondly, wind removes snow by physically overcoming the adhesive and cohesive forces or by shaking branches. Lastly, it increases snow load by delivering the snow at an angle and at force which allows the snow to penetrate the interior of complex canopies where it can be intercepted. Wind thus can increase or decrease the effective intercepting surface of complex structures.

The JGFES (1952) developed three models for testing the effect of wind on snow load. One was a series of equal-sized, hexagonal boards stacked one on top of another to simulate square canopies of varying complexity (see Fig. 7.4). Another was a series of hexagonal boards of varying sizes stacked at varying densities to simulate open, conical crowns of varying complexity (see Fig. 7.5). The third was a series of solid 'hexagonal pyramids' of varying height:base ratios to simulate closed crowns of varying shapes (see Fig. 7.25).

Figure 7.4 illustrates the effect of wind speed on snow load in modelled, rectangular, open crowns. As wind speed increased, more snow was captured in progressively more complex crowns. At low wind speeds, snow fell vertically and the upper levels of the equal-sized layers acted as umbrellas, shielding lower levels. At high wind speeds, the snow was delivered at an angle and could fill the lower levels.

Regression equations for stacks of different numbers of boards (computed from data of JGFES, 1952: 123) are:

$$\text{3 boards } Y = 33.1 + 26.5\ V \quad (r^2 = 0.48,\ P \leq 0.05) \tag{7.1}$$

$$\text{5 boards } Y = 11.1 + 20.2\ V \quad (r^2 = 0.67,\ P \leq 0.05) \tag{7.2}$$

$$\text{7 boards } Y = 6.9 + 20.2\ V \quad (r^2 = 0.76,\ P \leq 0.05) \tag{7.3}$$

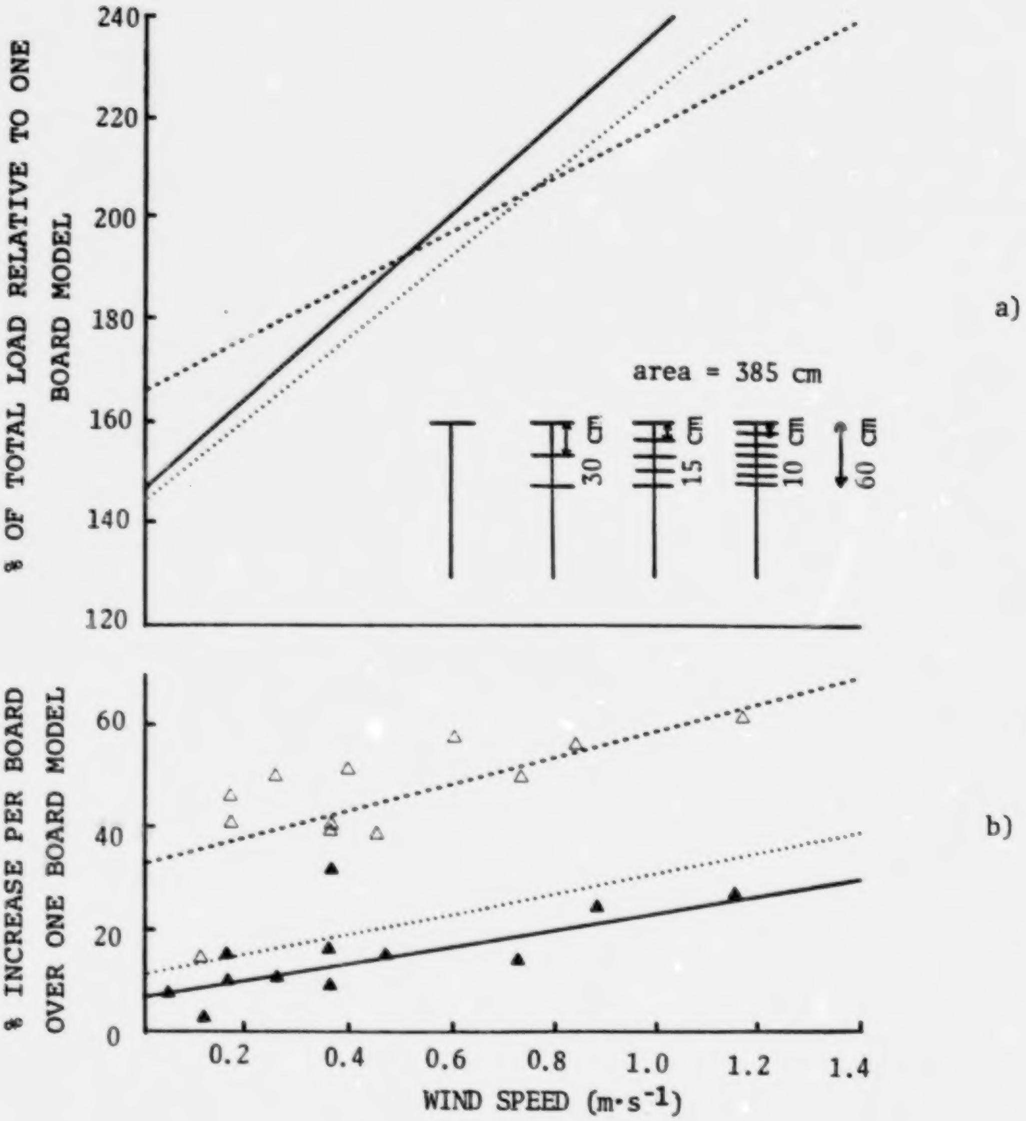


Figure 7.4 Effect of wind speed (m·s^{-1}) on the accumulation of snow in modeled rectangular, open crowns; 3 boards -----; 5 boards ·····; 7 boards ——.
a) The percent of snow load in models of different numbers of boards relative to the 1-board model.
b) The percent increase in snow load per board relative to the 1-board model; data points illustrated only for 3-board and 7-board models (data of JGFES 1952: 123).

where Y = the % increase in snow per board, over the amount of snow present on the one board model, and V = wind velocity in $m \cdot s^{-1}$ ($V < 1.5\ m \cdot s^{-1}$). With moderate statistical manipulation a summary equation can be derived:

$$Y(n) = 87.9\ N^{-1\ 45} + 56.4\ N^{-0\ 44}\ V \qquad (7.4)$$

where Y(n) is the rate of increase of snow mass in a modeled 'crown' of N boards (% over that of a single board, which is assumed broadly equivalent to level ground) and V is as before. Equation 7.4 was derived using all data and assuming an underlying linear form with velocity, but a curvilinear relationship with the number of boards. In its linear form ($Y = a + bV$, Eq. 7.4) the percentage increase in snow load is a negative power function of N, the number of boards; that is, $a = cN^{-}$. Coefficients of determination for the regressions relating the exponents of Eq. 7.4 to the number of boards were -0.998 and -0.996, respectively.

Because total height of the modeled crown remained the same (60 cm, Fig.7.4a), N is related to d (interval between boards, cm) by $N = 60/d$. Thus equation 7.4 can be expressed:

$$Y(n) = 0.235\ d^{1\ 45} + 9.384\ d^{0\ 44}\ V \qquad (7.5)$$

or $$Y(n) = 1/4\ d^{2\ 5}\ (d + 40\ V) \qquad (7.6)$$

where Y(n) and V are as before and d is the interval between boards (cm). The implication from the modeled crowns is that at moderate snowfalls with wind speeds (V) of 1 $m \cdot s^{-1}$, crowns with branch whorls 30 cm apart would intercept 75% of the snow falling on the level. Extrapolating well beyond the data, crowns with whorls 60 cm apart could retain up to 146% of the snow falling on the level. Were the wind speed 50% greater ($V = 1.5\ m \cdot s^{-1}$), the calculated values are 95% and 174%. While the specific magnitudes predicted are likely unrealistic for real trees and are extrapolated beyond the data, the form of the relationship is probably correct.

The effect of wind speed on snow load in modeled, conical open crowns likely approximates natural situations more closely (Fig. 7.5). The equations of Figure 7.5 are from the Japanese text (JGFES 1952: 131-132). Our computations using the data of their Table 3 (which match the 'scattergrams' of their Figure 9, but not the equations they applied to the Figure) yield different results:

Model 2 $Y = 105.9 - 30.1\ V$ $(r^2 = -0.19)$ (7.7)

Model 3 $Y = 88.5 - 1.6\ V$ $(r^2 = -0.003)$ (7.8)

Model 4 $Y = 107.3 + 13.13\ V$ $(r^2 = 0.13)$ (7.9)

Model 5 $Y = 111.4 + 29.6\ V$ $(r^2 = 0.41)$ (7.10)

where Y and V are as before (Eq. 7.1 through 7.6). Only the slope of model 5 (the most open configuration, Fig. 7.5a) is statistically different from zero using our calculations. As all the text was not translated we may have misinterpreted

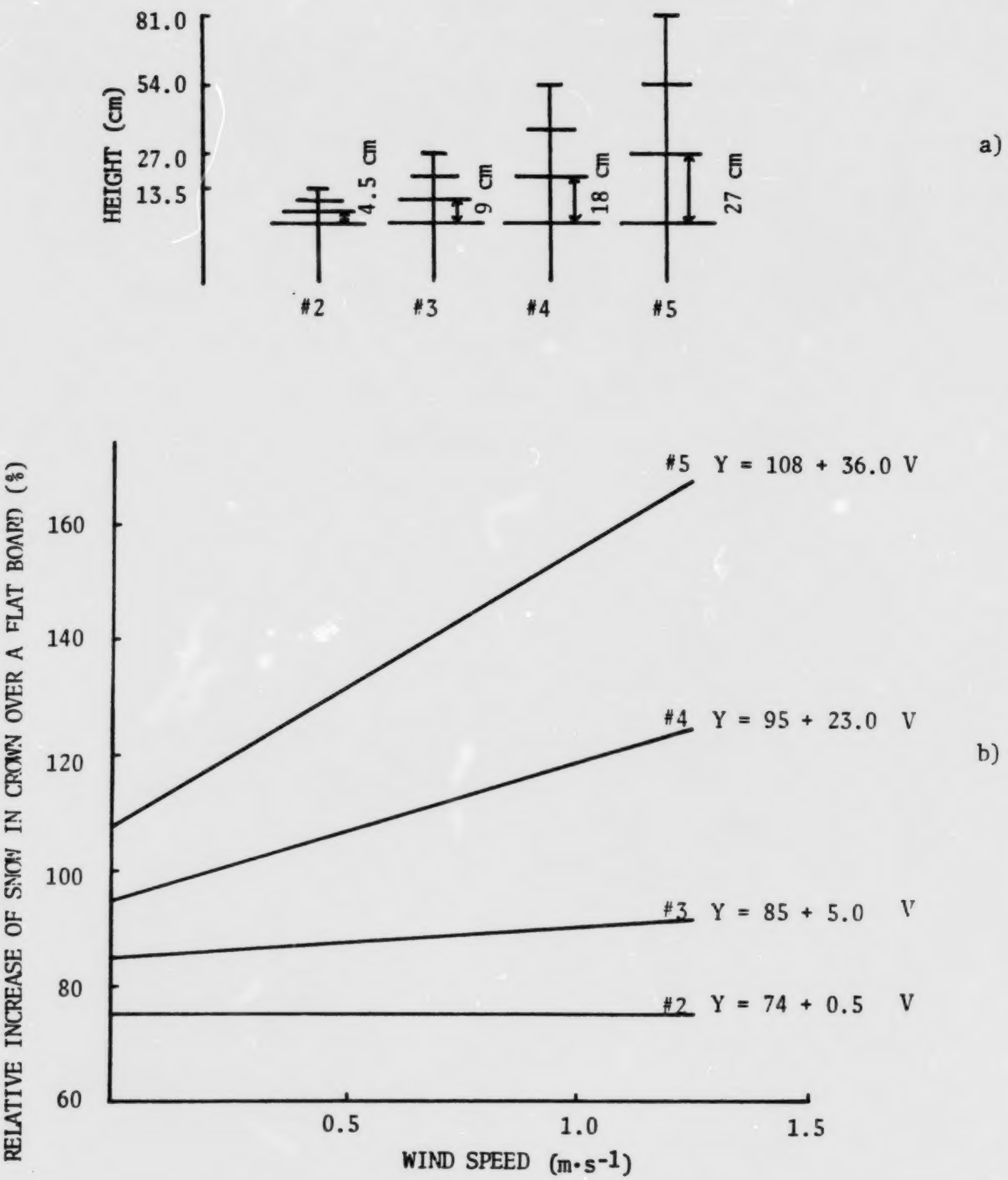


Figure 7.5 Effect of wind speed ($m \cdot s^{-1}$) on the accumulation of snow in modeled conical, open crowns.
a) Structure and dimensions of the models.
b) Relative increase of snow in modeled crowns over that of a flat board as a function of wind speed (data of JGFES 1952: 130).

something. The data as presented in their Table 3 indicate that at very dense configurations, increasing wind speed decreased snow load. In the more open cones (models 4 and 5), the snow load increased with increasing wind speed. The reduction in snow load of dense configurations (models 2 and 3 of Fig. 7.5a) with increasing wind speeds is greater using our calculations (Eq. 7.7 and 7.8) than JGFES calculations (Fig. 7.5).

We used the same approach to derive a summary equation as for the effect of number of boards (Eqs. 7.4 and 7.5). For any given model, the effect of wind speed on snow load was linear ($Y = a + bX$, Fig. 7.5). The regression coefficients, however, were related to distance between boards by a power function ($Y = aX$, Fig. 7.6). Because of the apparent ambiguity we present the summary equations both as calculated from the regressions as given in the Japanese text (Figs. 7.5 and 7.6) and as calculated from the data as presented by JGFES (1952: Table 3, p. 129). Best fits for the coefficients of the summary equation are:

JGFES regressions

$$a = 74d^{0\ 2} \qquad (7.11a)$$

$$b = d^{1\ 9} \qquad (7.12a)$$

JGFES data

$$a = 90.5d^{0\ 05} \qquad (7.11b)$$

$$b = 0.04d^{1\ 97} \qquad (7.12b)$$

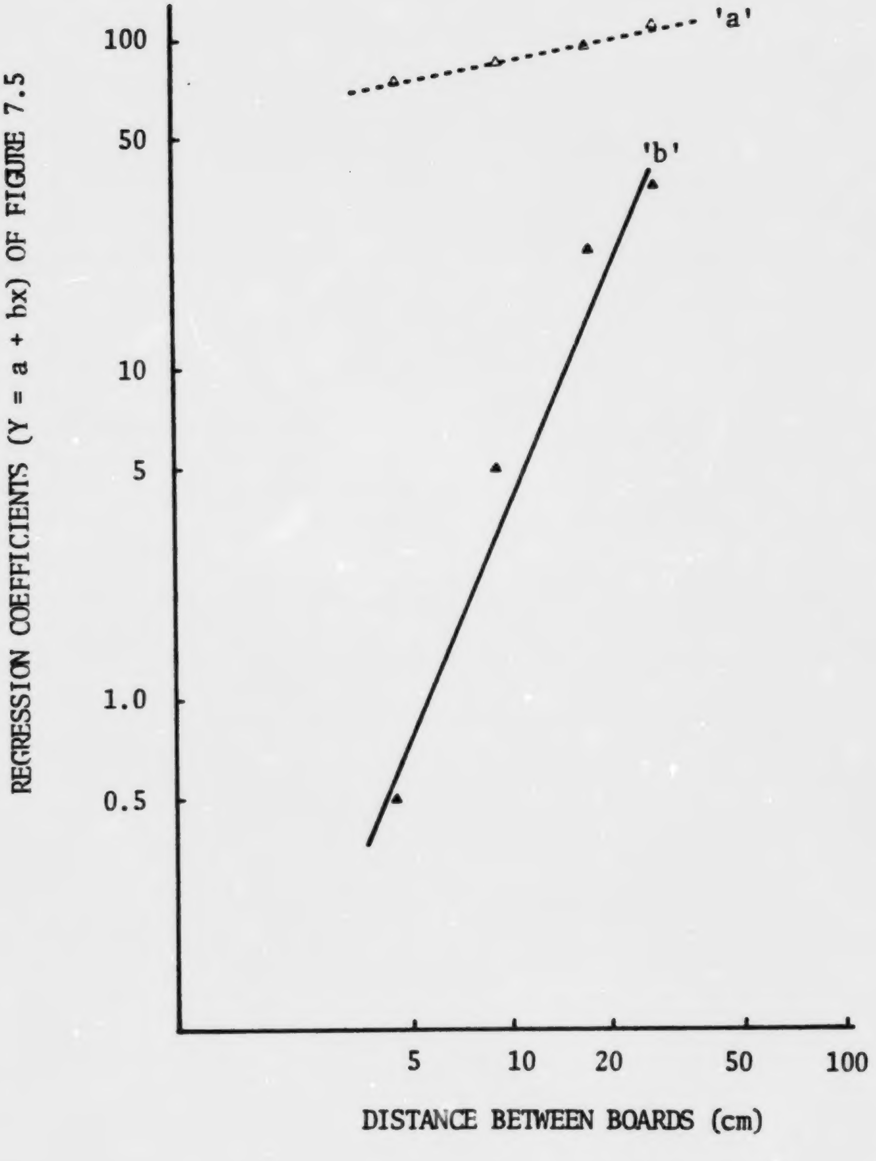


Figure 7.6 Coefficients of regression equations relating relative snow load in crown to that on the level, Y, as a function of wind speed, V (Y = a + bV) versus the distance between boards in the modeled crown. Regression coefficients from Figure 7.5; regression equations illustrated are Eqs. 7.11a and 7.12a.

where d is the distance between boards (Fig. 7.5).

The summary equation is thus:

JGFES regressions

$$Y(n) = 74d^{0.2} + d^{1.9}V \tag{7.13a}$$

JGFES data

$$Y(n) = 91d^{0.05} + 0.04d^{1.97}V \tag{7.13b}$$

where $Y(n)$ = % increase of snow accumulated on a model crown of n levels over that of a single board, d is the interval between levels (cm), and V is wind velocity ($m \cdot s^{-1}$). The empirical relationship holds only for conditions before the spaces between boards have filled up to assume a pyramidal shape.

JGFES summary equation 7.13a predicts an increase in snow load for model 4 ($d = 18$ cm) of 374.6% over a level board for a wind speed of 1 $m \cdot s^{-1}$. The observed values were 106 and 126% at wind speeds of 0.72 and 1.16 $m \cdot s^{-1}$ respectively. Our equation (7.13b) predicts an increase of 117% for the same conditions. It more accurately reflects the data as provided and suggests a strong effect of inter-whorl distance on snow interception. The relationship with velocity is again linear once the effects of inter-whorl distance have been incorporated.

We interpret the observation that short inter-whorl distances reduce snow load at increasing wind speeds to result from a 'slope effect': dense configurations intercepting snow assume a conical shape and readily shed snow from their sloping slides. The interpretation implies that with increasing wind speed, slope plays an increasing role in reducing snow load. Whereas snow load increases with wind speed in model trees of level surfaces (Figs. 7.4 and 7.5), it should decrease with increasing wind speed in trees with steeply sloping surfaces. The interpretation appears correct and is corroborated by research on model surfaces (Figs. 7.24 and 7.25) and actual tree crowns (Figs. 7.7 and 7.10).

The crown of Cryptomeria is pyramidal or steeply sloping. Data in Figure 7.7 indicate that although snow load was variable (changing temperature effects), there was a general tendency for snow load to decrease with increasing wind speed. Compared to the deposit of snow at zero wind, that at 1 $m \cdot s^{-1}$ was decreased about 25 to 30%. At higher wind speeds the Cryptomeria crown acted like a solid cone.

Takahashi (1953) reported that intercepted snow on the sloping branches of Cryptomeria did not accumulate at wind speeds greater than 3.5 $m \cdot s^{-1}$. That observation is in agreement with Figure 7.7. The complex interaction of slope and wind speed is treated more fully in the discussion of effects of individual tree morphology on snow delivery (Ch. 7.6).

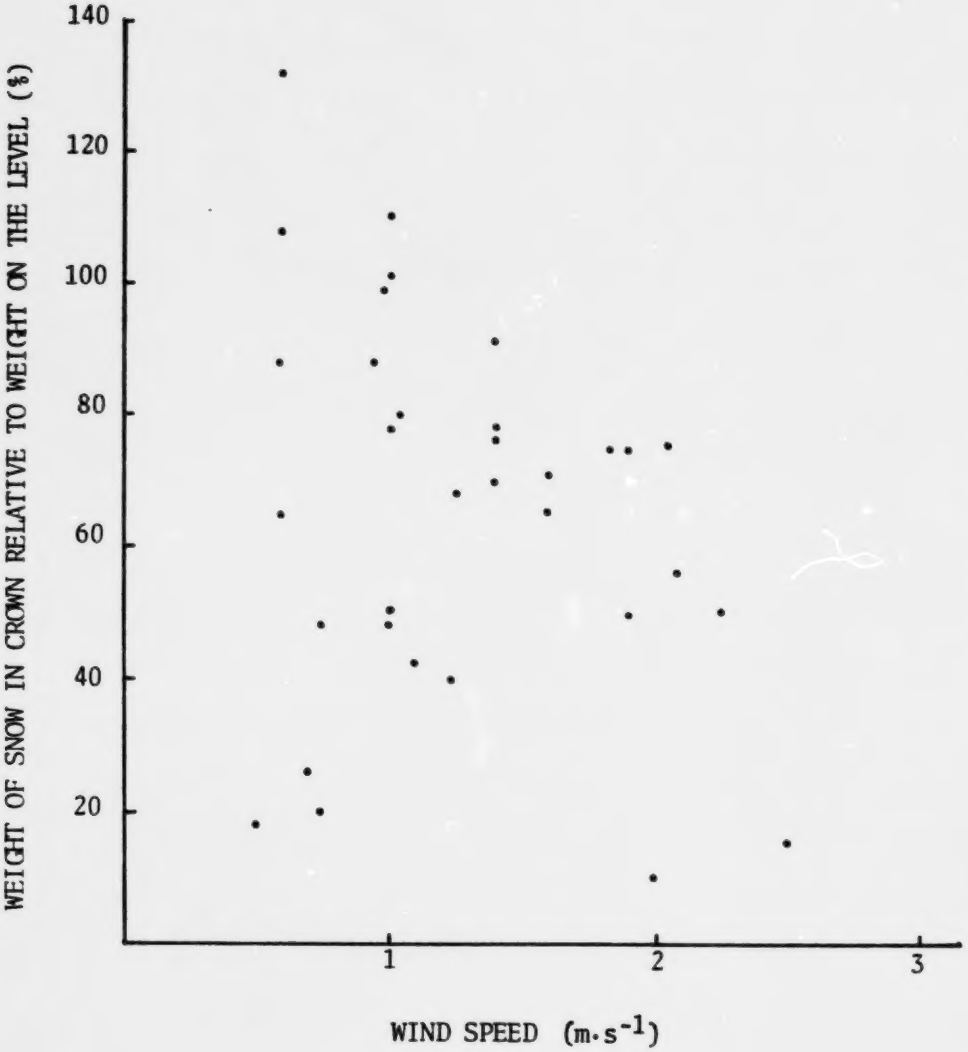


Figure 7.7 The effect of wind speed on weight of snow in the crown of a Cryptomeria tree during 34 storms. This is the same tree that was illustrated in Figure 7.2 (data from JGFES 1952: 145).

7.4 Temperature-Wind Interactions

Whether snow delivered to a potential intercepting surface during a storm stays on that surface, is a function of both wind and temperature. The potential for wind to redistribute snow is affected by cohesion among snow particles and adhesion to their supporting surfaces. Both are functions of temperature. Cold, dry snow crystals do not tend to pack or gain strong cohesive forces. Cohesive forces are intensified at temperatures just below 0°C. Warmer intercepting surfaces tend also to increase adhesive forces (Ch. 7.1). One result is that cold, dry snow is blown from trees more easily than warm, sticky snow (Miller 1964). Adhesion is reduced significantly, however, when temperatures rise enough to cause melt at the boundary between snow and its intercepting surface (Ch. 7.2, Fig. 7.10). Refreezing of the boundary layer increases adhesion (Miller 1966). The influence of temperature on adhesion and cohesion is summarized in Figure 7.8. Wind itself increases cohesion and adhesion by increasing the impact velocity of snow crystals.

The roughness of the intercepting surface has an additional effect upon adhesion, with rougher surfaces acquiring more snow. In southcoastal British Columbia that would result primarily because rougher surfaces can retain more liquid water (Horton 1919, Delfs 1955). Furthermore, depending on the flexibility of the surface, wind may supply sufficient energy to cause widespread breakdown between the

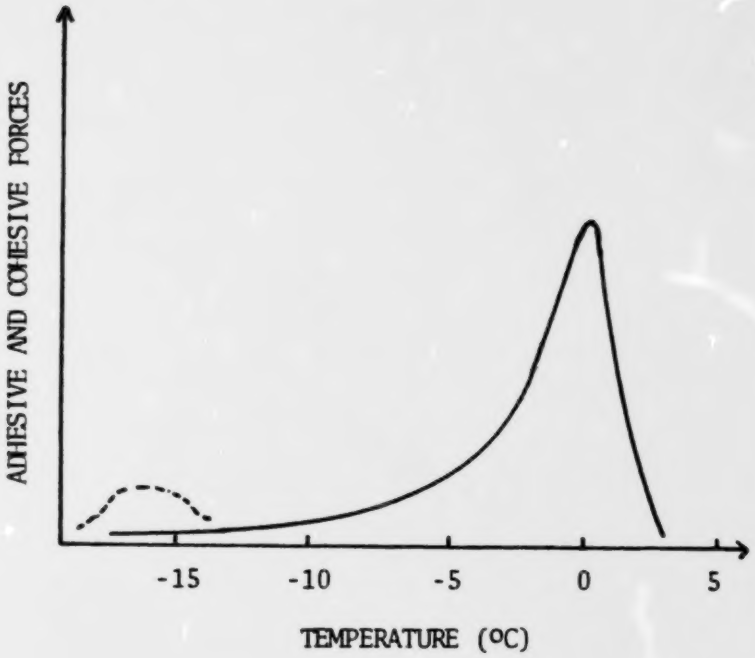


Figure 7.8 Conceptual model depicting the effect of temperature on adhesive and cohesive forces between an intercepting surface and snow. The dashed line indicates increased cohesion at temperatures encouraging dendritic, interlocking crystals.

snow load and the intercepting surface; especially if the temperature is warm (Fig. 7.10). As a result, there is wide variation in the literature concerning wind's ability to counteract the physical forces holding snow onto the intercepting surface (Table 7.1). The relatively large scale break down of adhesion usually results in large clumps of snow being knocked off the intercepting surface. Depending on the cohesive forces within the clump it may break up sufficiently for parts to be carried by the wind or it may fall intact to the snowpack directly below the intercepting surface. Miller (1962) reported that in forests these snow masses are too dense to be blown far. As a result they fall to the snowpack below the canopy and create an increase in density as well as an increase in depth over the area of impact. However, Miller's conclusion is controversial; other workers (e.g., Hoover and Leaf 1967) concluded that snow intercepted by trees is often redistributed by wind to forest openings (see also Ch. 8.3). Presumably the different interpretations are a result of different temperature conditions. In warmer, more maritime areas, snow is less likely to be redistributed by wind.

Broadly, the data reviewed indicate that, particularly at temperatures close to freezing, moderate increases in wind speed can increase the snow load whenever low angles of delivery allow the snow to penetrate complex canopies. The diagonal vector of the snow effectively increases the intercepting surface, especially if component surfaces are

Table 7.1 Published observations on the effects of wind on snow removal from trees.

Source	Wind speed or influence	Location	Intercepting surface	Comments
JGFES (1952)	increase from calm to 1 $m \cdot s^{-1}$	Japan	*Cryptomeria* tree	25% reduction in snow load
JGFES (1952)	calm to 3 $m \cdot s^{-1}$	Japan	*Cryptomeria* tree	50% reduction in snow load
Hoover and Leaf (1967)	light physical shaking of intercepting surface	Colorado	*Pinus contorta* *Picea engelmannii*	almost 100% reduction in snow load (≤ 30°F; -1.1°C)
Pruitt (1958)	moderately strong winds	Alaska	*Picea* tree	very little reduction in snow load
Cramer (1960)	blast of wind from helicopter		*Picea* (?)	only top 2-3 metres of trees affected
Sakharov (1949)	?		*Picea* trees	snow blows from upper storey but not from sheltered regeneration
Goodell (1959)	violent shaking to simulate strong wind	Colorado	*Picea engelmanni tree (4.5 m)*	1/3 of fresh dry snow could be shaken out of tree

horizontal (Figs. 7.4 and 7.5). Naturally, the angle of intercepting branches becomes more acute with increasing snow load (Ch. 7.6). On sloping surfaces wind speeds decrease the snow load with the effect being greater for steeper slopes (Fig. 7.24).

One apparent effect of the temperature:wind interaction is that snow loads tend to be greater at night. Shidei (1954) noted that "the snow crown of the tree grows large and heavy at night when the air temperature falls to about 0°C and the air is calm with no [incoming] radiation". The phenomenon is also evident in data of JGFES (1952). Figure 7.9 illustrates the snow load for seven storms as measured on a continuously weighed Cryptomeria tree. Irrespective of the rate of snowfall, snowfall at night added more to the load on the tree than did snowfall by day. The most rapid accumulation occurred after 18:00 with peak loads being reached near sunrise (Fig. 7.9). Temperature data in Figure 7.9 unfortunately are not for the same days (usually 1 day different; JGFES 1952: 142, 143). One interpretation is that during calm night air at temperatures near -1°C, at which adhesive and cohesive are strongest, interception of snow occurs effectively. The snow load accumulates rapidly through night during calm weather; near sunrise both insolation and wind speeds increase and both adhesive and cohesive forces are less effective. With increase in insolation and wind speed the accumulated snow load drops from the crown (Fig. 7.10).

During daytime snow load may be decreased by insolation

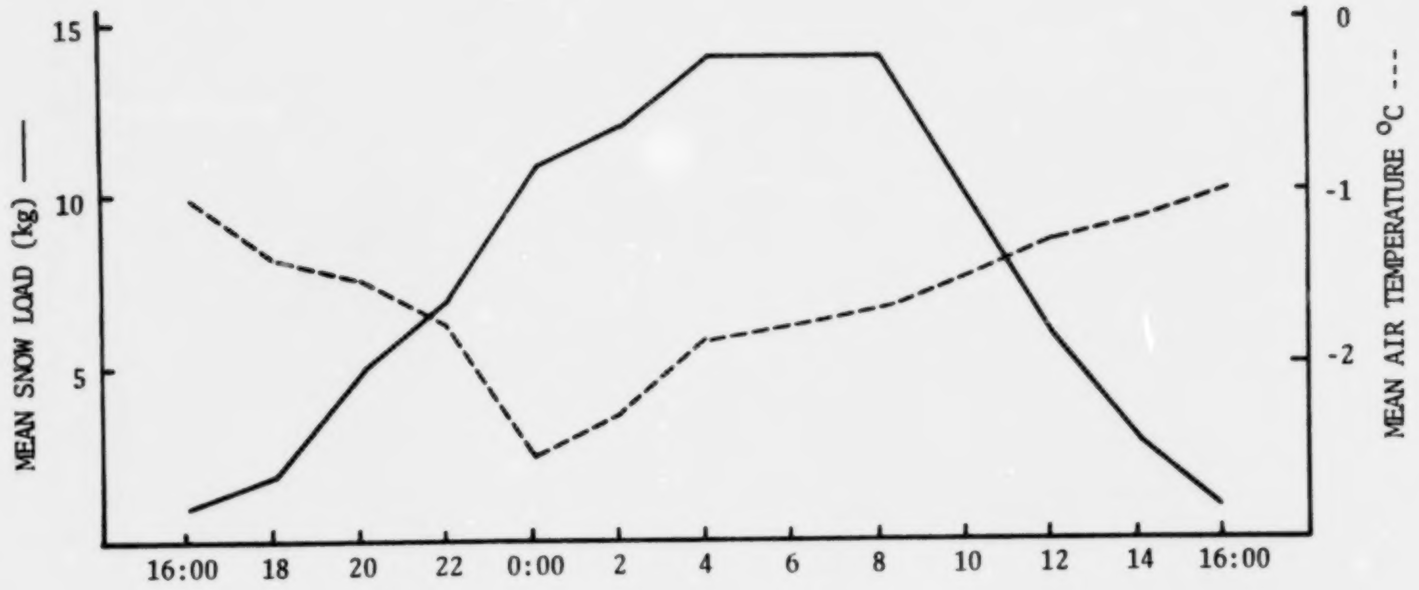


Figure 7.9 Mean diurnal pattern of snow load in a Cryptomeria crown during seven snow storms (data of JGFES 1952: 142,143).

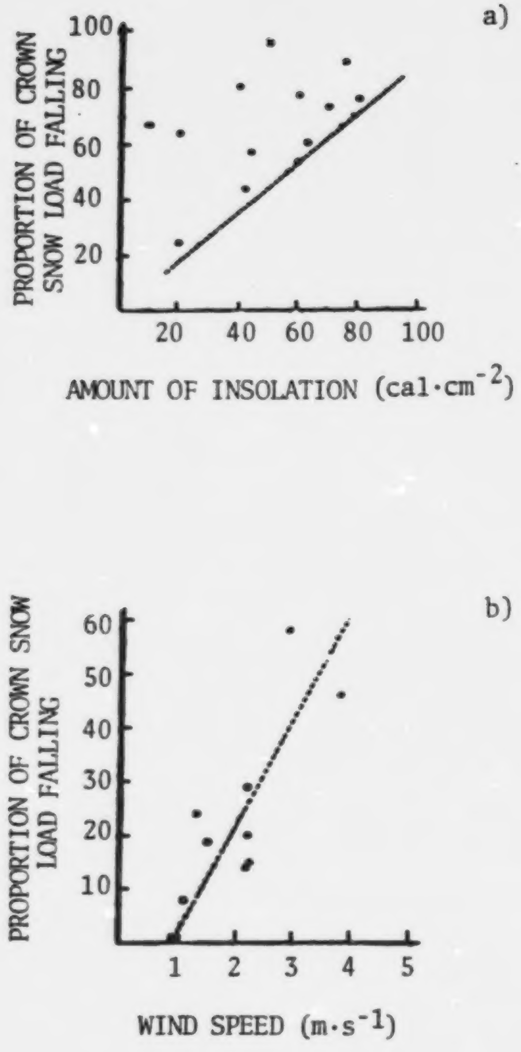


Figure 7.10 The proportion of snow falling from a Cryptomeria crown as a function of: a) amount of insolation and b) wind speed. Data are for cases when air temperatures were below freezing (from JGFES 1952: 146).

which provides some heat even while snow is falling. Figure 7.10a compiles data for 13 storms; the lower envelope including all storms has a slope of about 0.9% loss of snow load for each langley ($\mathrm{cal \cdot cm^{-2}}$) of insolation. With low radiation the rate of loss of intercepted snow could be slow or fast depending on air temperature; at high radiation it would always be fast (air temperature was below freezing for all storms in Fig. 7.10). If a crown carried a snow load of 8 mm SWE and continued to receive new snow at 1 $\mathrm{mm \cdot h^{-1}}$, diffuse incident radiation of 10 $\mathrm{ly \cdot h^{-1}}$ would generate an hourly rate of net loss of 9% of the load. By making plausible assumptions of the nature of short wave and long wave radiation and using the data illustrated in Fig. 7.10a, Miller (1966) calculated the radiative cost of mass transport of snow to be 1.5 ly for the release of 1 mm SWE. That calculation suggests that the release of intercepted snow occurs after about 20% of it has melted. Miller (1966: 7) used other data extracted from JGFES (1952) to calculate rates of mass transport at air temperatures above freezing. The most rapid rate (3 $\mathrm{mm \cdot h^{-1}}$) occurred at temperatures slightly above freezing when light snow was still falling; initial snow load was 14 mm SWE.

The broad interactions of wind speed and temperature effects on snow interception are summarized in Figure 7.11. Stated briefly they are:

1) In calm weather, the relation of snow load to temperature

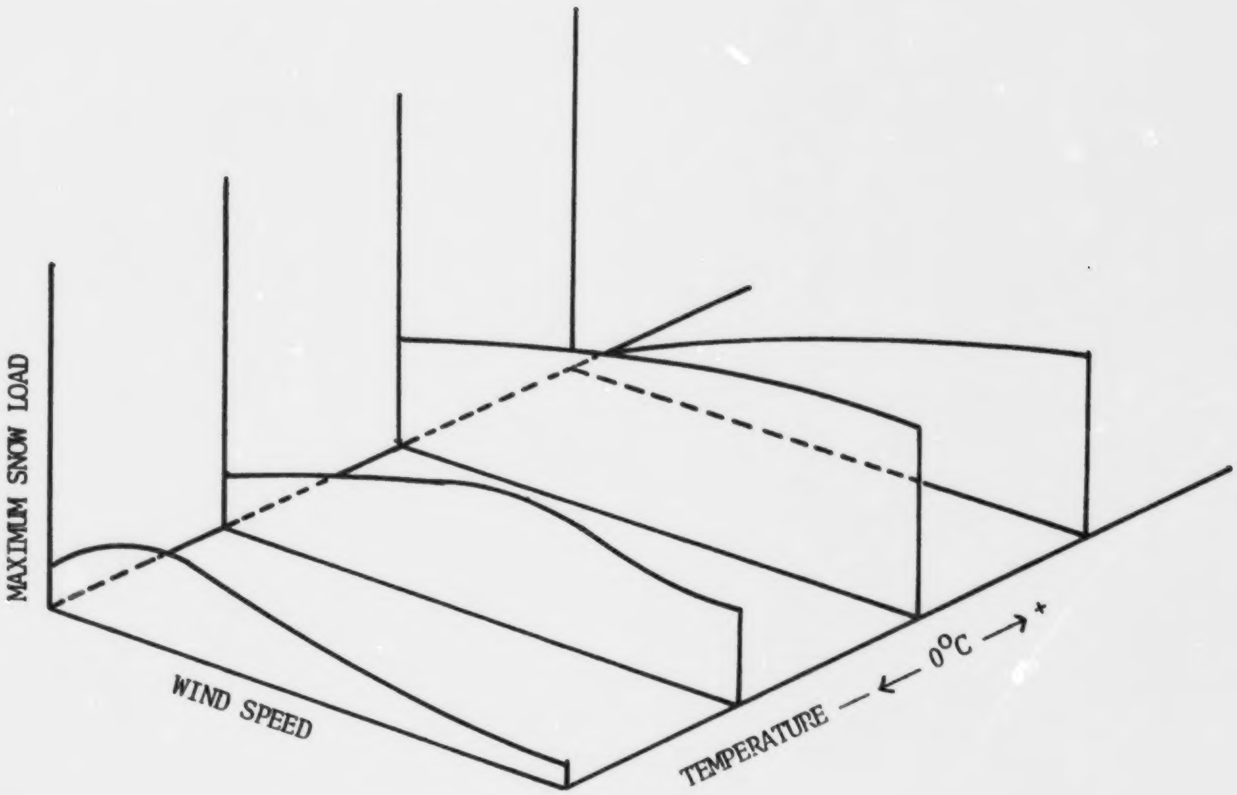

MAXIMUM SNOW LOAD
WIND SPEED
TEMPERATURE — ← 0°C → +

Figure 7.11 Schematic representation of the interaction of wind and temperature effects on snow interception (adapted from Miller 1964: Fig. 1).

is curvilinear with a maximum just below freezing (Fig. 7.2).

2) In cold storms, the relation of snow load to wind speed is curvilinear, with a maximum near 1 to 2 $m \cdot s^{-1}$. Snow load decreases rapidly at higher wind speeds, particularly on sloping surfaces (Fig. 7.7).

3) An increase of temperature up towards freezing lessens the effect of wind in reducing snow load.

4) In storms with temperatures just below freezing, snow load increases at low wind speeds and may exceed that generated in calm weather.

7.5 Effects of Storm Size

One factor neglected by many workers is that the interception efficiency of the canopy varies between storms of different sizes and at various times within the same storm (e.g., Fig. 7.9). The first few flakes of a snow storm encounter only the resistance to throughfall offered by the canopy density and adhesive forces in effect. As the storm continues and snow begins to accumulate in the canopy, effective intercepting surfaces are increased by the snow accumulation itself and the tendency of accumulating snow to push branches together forming interlocking platforms.

Adhesive forces give way to cohesive forces. The efficiency of interception increases throughout the storm until the maximal snow load is reached. Further snowfall drops through as overload throughfall. An additional complication is that snow load can modify the angles of the intercepting surfaces thus reducing their effectiveness, particularly under windy conditions (Ch. 7.6). When these individual processes are combined the curve representing intercepted snow against cumulative snowfall within a storm is sigmoid (Fig. 7.12).

The best data illustrating this process come from the JGFES (1952), Watanabe and Ozeki (1964), and Satterlund and Haupt (1967); these studies continuously weighed single trees in single snow storms. The expected sigmoid shape was evident in two of the studies (Figs. 7.13 and 7.14). The initial part of the curve was very nearly identical in both studies where it was documented, with a rapid change occurring at cumulative precipitation measurements of about 4 mm SWE or 4 $kg \cdot m^{-2}$. That is a small amount of precipitation and probably within the region of measurement error for many studies; particularly those comparing precipitation in the open and under the canopy. The upper asymptote of maximal snow load differed significantly between studies; 12-14 mm [$kg \cdot m^{-2}$] for JGFES (1952) (at about 18 mm precipitation) compared to 2.5-3.5 mm in Satterlund and Haupt's study (at about 7.5 mm precipitation). Sufficient data do not exist for us to evaluate potential reasons for this large difference in maximal snow load. The pine weighed by Satterlund and Haupt

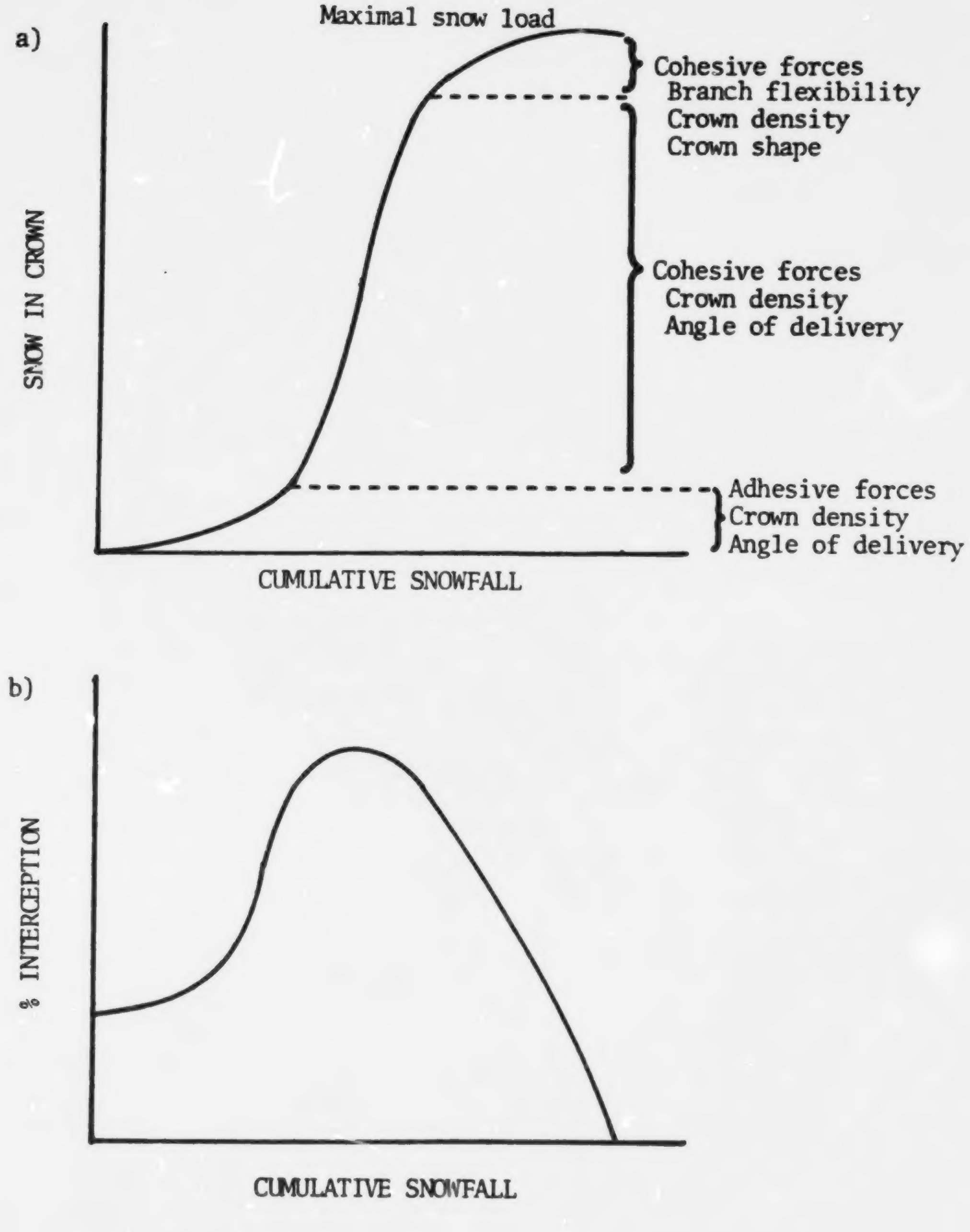


Figure 7.12 Schematic representation of the temporal pattern of: a) crown snow load, and b) % interception, as a function of cumulative snowfall.

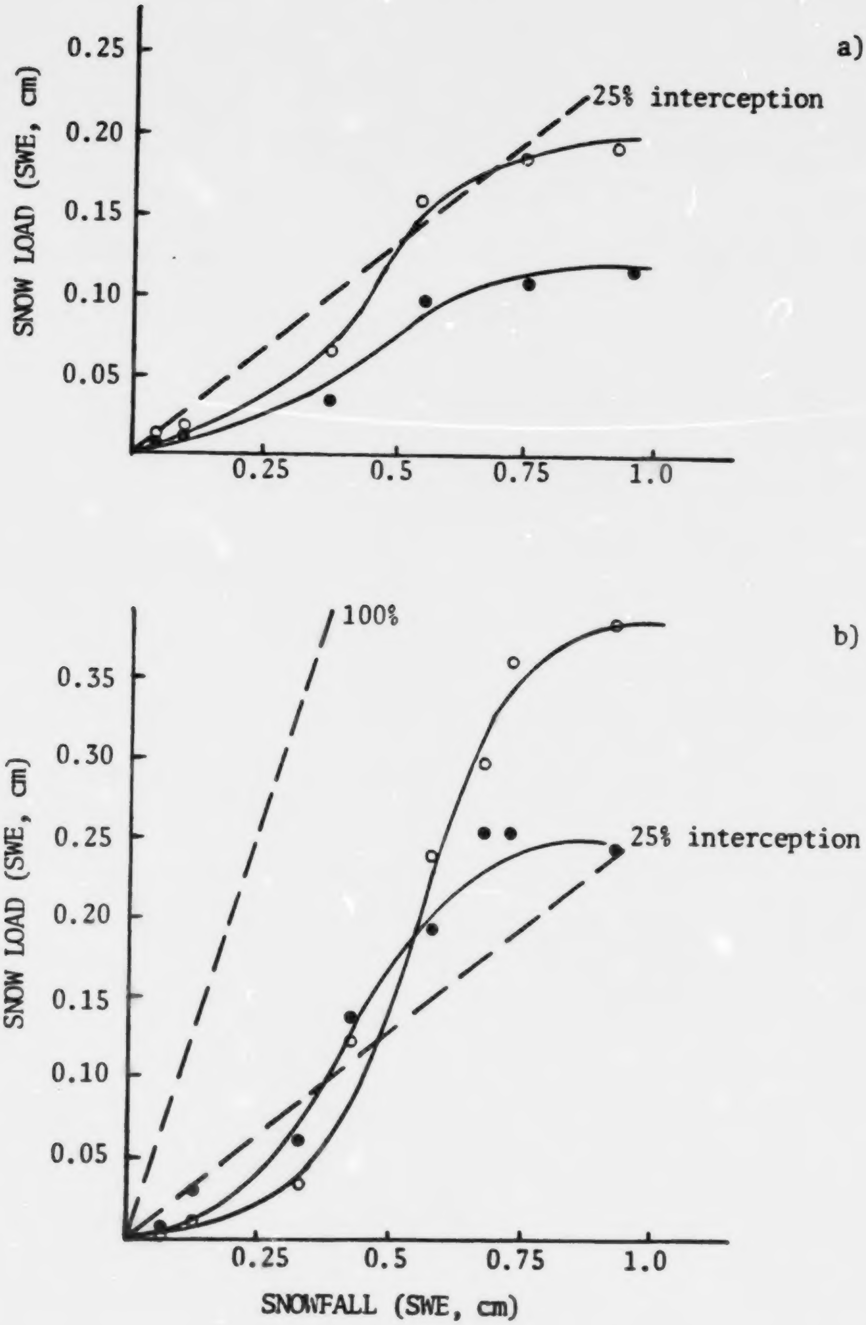


Figure 7.13 Snow catch by Douglas fir (o) and white pine (●) trees during two storms on January 10, 1967 (a) and January 12, 1967 (b) (data of Satterlund and Haupt 1967: 1038).

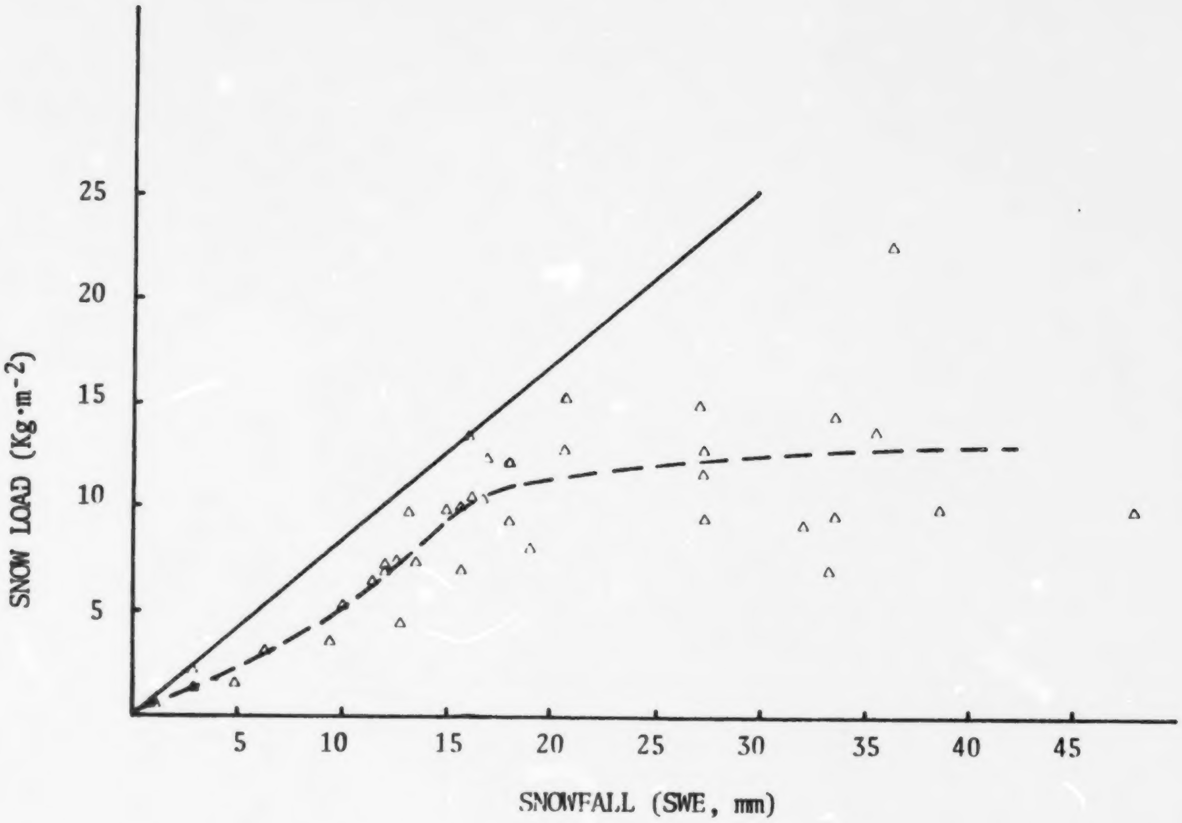


Figure 7.14 Snow catch by a *Cryptomeria* tree weighed continuously during several storms (-----). The solid line represents the maximum envelope with an interception efficiency of about 85% (data of JGFES 1952: 142).

(1967) was 3.7 m tall, the Douglas-fir was 4.2 m. Photographs of the *Cryptomeria* suggest it was considerably taller. Asymptotes for the two varieties of *Cryptomeria* studied by Watanabe and Ozeki (1964) were about 15 and 25 mm SWE at 80 mm of precipitation (Fig. 7.15). Crown depths of their study trees ranged from 2.4 to 6.1 m averaging near 5 m. The magnitude but not the pattern of interception differed among the three studies indicating that the conceptual model of Figure 7.12a is broadly correct.

Table 7.2 summarizes published data on snow loads held by single trees. Only *Eucalyptus* trees in Australia (Costin et al. 1961) held snow loads in excess of 7 mm SWE. *Eucalyptus pauciflora* has heavy leaves and stiff twigs and petioles which mat down, as snow accumulates, to form a platform capable of bearing a heavy load (see also Table 7.4). The branches apparently are rigid, maintaining horizontal platforms such as those of the model trees in Figure 7.4.

Heavy snow loads also could result from extremely wet snow. Klintsov (1958) reported gales of wet snow in Sakhalin during which snow accumulated in crowns of conifers about 7 m tall to nearly 100 kg. Similarly, Heikenheimo (1920) reported wet storms in Finland in which rime and snow were plastered on trees by high winds to depths of 30 cm. The moisture encouraged adhesion and cohesion.

Because the conceptual model (Fig. 7.12a) appears to mimic empirical observations, the hypothetical pattern of interception efficiency during a storm also should hold

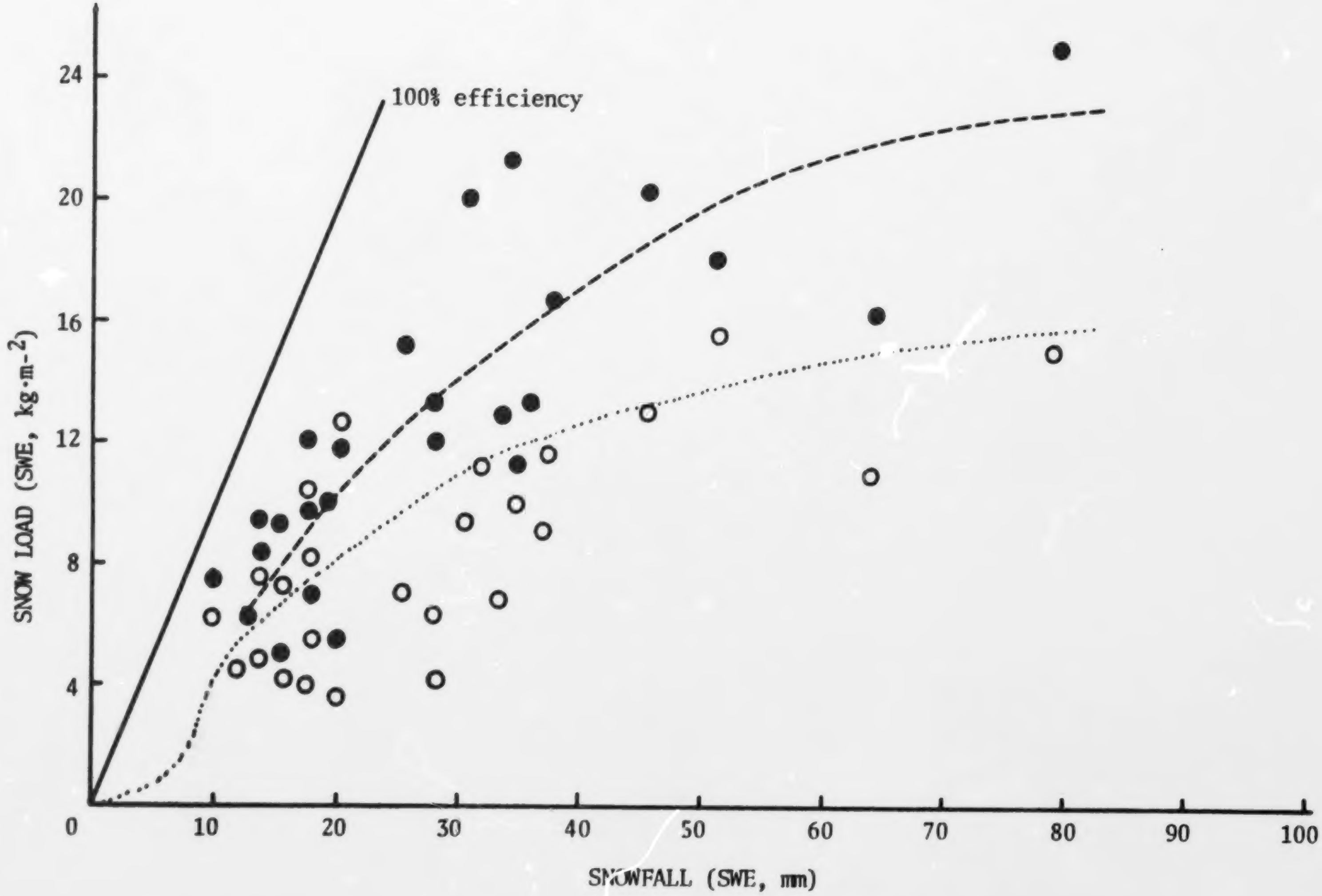


Figure 7.15 Snow catch by two varieties of Cryptomeria (● = Bokasugi; o = Kumasugi, snow adapted) during one continuous snowfall. The hypothetical form of initial part of the curve is illustrated for Kumasugi (data from Watanabe and Ozeki 1964: 125).

Table 7.2 Published studies of snow interception by single trees.

Source and Location	Tree species	Weather	Cumulative amounts of snow: Intercepted	Cumulative amounts of snow: Fallen	% intercepted	Remarks
Satterlund and Haupt (1967) Idaho	Pseudotsuga menziesii (13.7 ft high; 10.68 ft crown diameter; 89.58 ft² crown area; 514.61 ft² needle area)	0 to 0.6°C	0.13 mm SWE	0.51 mm SWE	25	Trees were continuously weighed during two snowstorms. Interception was described by a sigmoid growth curve (Fig. 7.13).
			0.20 mm SWE	1.02 mm SWE	20	
			0.64 mm SWE	3.81 mm SWE	17	
			1.52 mm SWE	5.59 mm SWE	27	
			1.80 mm SWE	7.62 mm SWE	24	
			1.88 mm SWE	9.40 mm SWE	20	
		-0.6 to 0.6°C	0.00 mm SWE	0.51 mm SWE	0	
			0.10 mm SWE	1.27 mm SWE	8	
			0.30 mm SWE	3.30 mm SWE	9	
			1.19 mm SWE	4.32 mm SWE	28	
			2.29 mm SWE	5.84 mm SWE	39	
			2.90 mm SWE	6.86 mm SWE	42	
			3.48 mm SWE	7.37 mm SWE	47	
			3.71 mm SWE	9.40 mm SWE	39	
	Pinus monticola (12.0 ft high; 10 ft crown diameter; 78.54 ft² crown area; 95.31 ft² needle area)	0 to 0.6°C	0.10 mm SWE	0.51 mm SWE	20	
			0.13 mm SWE	1.02 mm SWE	13	
			0.33 mm SWE	3.81 mm SWE	9	
			0.97 mm SWE	5.59 mm SWE	17	
			1.09 mm SWE	7.62 mm SWE	14	
			1.17 mm SWE	9.40 mm SWE	12	
		-0.6 to 0.6°C	0.03 mm SWE	0.51 mm SWE	5	
			0.30 mm SWE	1.27 mm SWE	24	
			0.58 mm SWE	3.30 mm SWE	18	
			1.35 mm SWE	4.32 mm SWE	31	
			1.91 mm SWE	5.84 mm SWE	33	
			2.46 mm SWE	6.86 mm SWE	36	
			2.46 mm SWE	7.37 mm SWE	33	
			2.39 mm SWE	9.40 mm SWE	25	
Goodell (1959) Colorado	Pica engelmannii (4.5 m high)	Cold, dry snow well below 0°C	16 kg per tree (ca 5 mm SWE)	1 cm SWE	ca 50	Weighed before and after storms.

Table 7.2 (continued)

Miller (1964)	Pinus contorta (young)	?	6.1 mm SWE	?	?	Snow beaten off and weighed.
Costin et al. (1961) Australia	Eucalyptus pauciflora	?	100 ft^3 per tree or 25 mm SWE	?	?	
Lull and Rushmore (1961) New York	'Balsam Fir'	?	5.41 mm SWE			Branches were nailed to a post and snow depth measured. Data here are recalculated assuming density of 0.1g·cm-^{3}
	'Red spruce'		4.67 mm SWE			
	'White pine'		4.52 mm SWE			
	'Hemlock'		3.00 mm SWE			
JGFES (1952)	Cryptomeria	-4 to +2°C				Sigmoid form illustrated in Figure 7.14.
Watanabe and Ozeki (1964) Japan	Cryptomeria japonica (variety kamasugi)		9.9 mm SWE	32.1 mm SWE	31	12 independent storms, trees weighed before and after storm.
			11.5 mm SWE	34.3 mm SWE	34	
			22.3 mm SWE	165.6 mm SWE	13	
			5.0 mm SWE	13.9 mm SWE	36	
			18.4 mm SWE	38.9 mm SWE	47	
			7.4 mm SWE	27.7 mm SWE	25	
			10.9 mm SWE	67.1 mm SWE	16	
			3.7 mm SWE	19.5 mm SWE	19	
			6.6 mm SWE	29.7 mm SWE	22	
			9.6 mm SWE	36.4 mm SWE	26	
			6.8 mm SWE	34.8 mm SWE	20	
			4.2 mm SWE	17.7 mm SWE	24	
	Cryptomeria japonica (variety bokasugi)		20.3 mm SWE	32.1 mm SWE	63	12 independent storms, trees weighed before and after storms.
			21.5 mm SWE	34.3 mm SWE	63	
			31.9 mm SWE	165.6 mm SWE	19	
			8.9 mm SWE	13.9 mm SWE	64	
			34.0 mm SWE	38.9 mm SWE	87	
			15.5 mm SWE	27.7 mm SWE	56	
			16.7 mm SWE	67.1 mm SWE	25	
			5.7 mm SWE	19.5 mm SWE	29	
			13.7 mm SWE	29.7 mm SWE	46	
			13.7 mm SWE	36.4 mm SWE	38	
			13.3 mm SWE	34.8 mm SWE	38	
			7.3 mm SWE	17.7 mm SWE	41	

(Fig. 7.12b). Data of Satterlund and Haupt (1967) in Table 7.2, show variability in the pattern between storms (Figs. 7.13 and 7.16). Both storms illustrated occurred in calm weather and both produced the same amount of snowfall (9.4 mm); that of January 10 lasted 6 h, that of January 12 lasted 8 h. Temperatures of January 12 began at 33°F [0.6°C] and declined to 32°F [0°C] in 1 h and 31°F [-0.6°C] in 4 h. The temperature pattern apparently encouraged rapid adhesion and cohesion once freezing occurred; interception efficiency increased dramatically and the hypothetical pattern of Figure 7.12b was exhibited (Fig. 7.16). In the more intensive storm of January 10, temperatures began at 0°C and stayed there until the last hour (0.6°C). Interception efficiencies began at higher levels than on January 12, but then declined (Table 7.2, Fig. 7.16). It is apparent that the conceptual model of Figure 7.12 can be modified by minor changes in temperature or precipitation intensity. Cohesion apparently was not as effective in the storm of January 10 (Fig. 7.13).

Following the same rationale as that used to generate Figure 7.12 we would expect maximum snow loads for given storms to exhibit a similar pattern with total snowfall as does snow load during a storm with cumulative snowfall. Figure 7.17 illustrates the initial part of the curve for two varieties of Cryptomeria; the expected exponential relationship is evident. If overload throughfall occurs we would expect total snow loads to decline at much greater snowfalls, and they do. At snowfalls of 67.1 and 165.6 mm

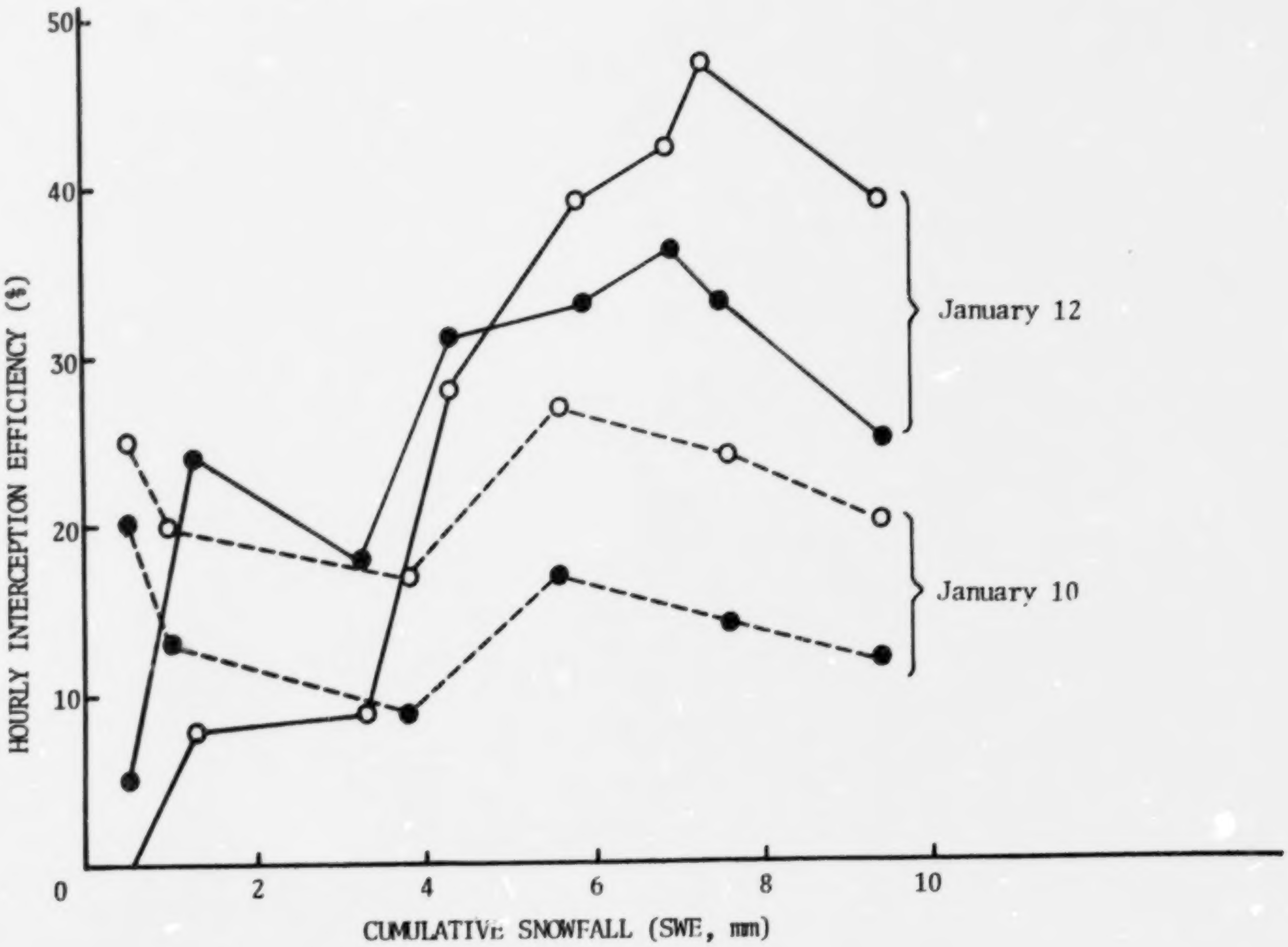


Figure 7.16 Interception efficiency of Douglas fir (o) and western white pine (●) as a function of cumulative snowfall during two different storms (data of Satterlund and Haupt 1967: Tables 2 and 3).

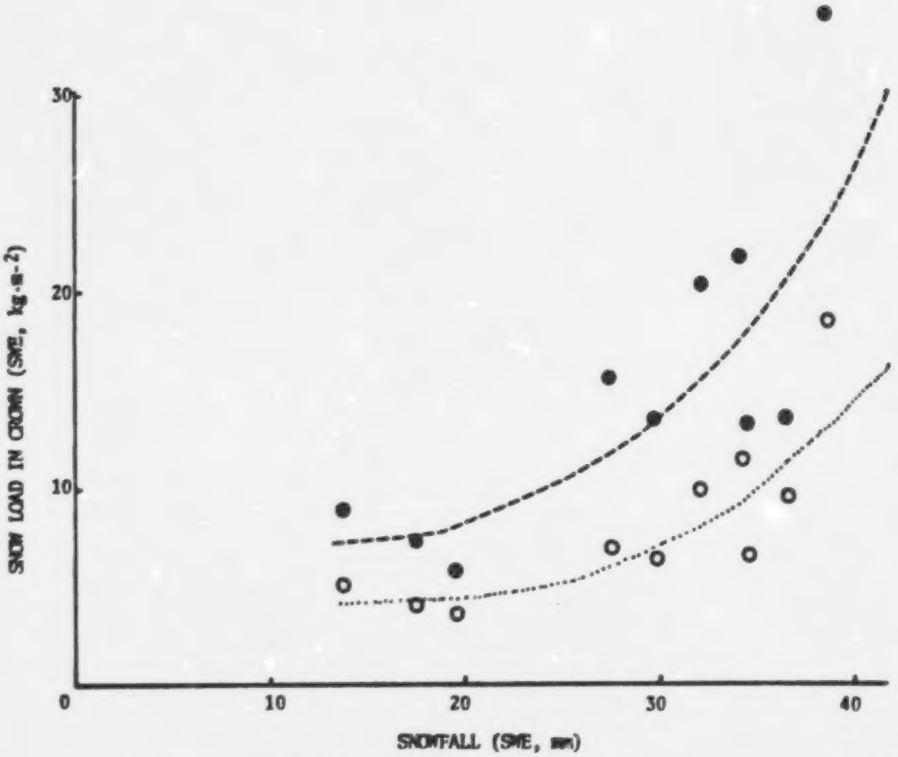


Figure 7.17 Maximum snow load in crowns of two varieties of *Cryptomeria* (● = Bokasugi, o = Kumasugi; snow adapted) as a function of total snowfall during different storms. Data are for individual trees in separate storms (data from Watanabe and Ozeki 1964: Table 3).

total snow loads declined to 10.9 and 16.7 and 22.3 and 31.9 $kg \cdot m^{-2}$ SWE for Kumasugi and Bokasugi varieties of Cryptomeria, respectively. We also expect the pattern of interception efficiency for maximum snow load among individual storms of varying size to approximate that of Figure 7.12b. The expectations are again broadly satisfied (cf. Figs. 7.12b and 7.18). The conceptual model for interception by individual trees appears to hold. A considerable portion of the variability in observed values, and their small scale departure from the conceptual model, likely arises because of loss of intercepted snow during a storm (e.g., Fig. 7.10). The model does not incorporate such losses because these are temperature, radiation, and wind dependent, thus erratic.

7.6 Effects of Tree Morphology on Snow Interception

The set of processes generating the basic pattern of interception by trees (Fig. 7.12) does not differ between trees of different morphologies (e.g., Harestad and Bunnell 1981). Thus both varieties of Cryptomeria (Figs. 7.15, 7.17, 7.18) or both Douglas fir and western white pine (Figs. 7.13 and 7.16) show similar patterns of interception versus snowfall under the same environmental conditions. However, the magnitude of those patterns can differ markedly. That is best illustrated by the storm of January 10 where the pattern of interception is identical but the magnitude differs markedly between Douglas-fir and white pine (Figs. 7.13a and

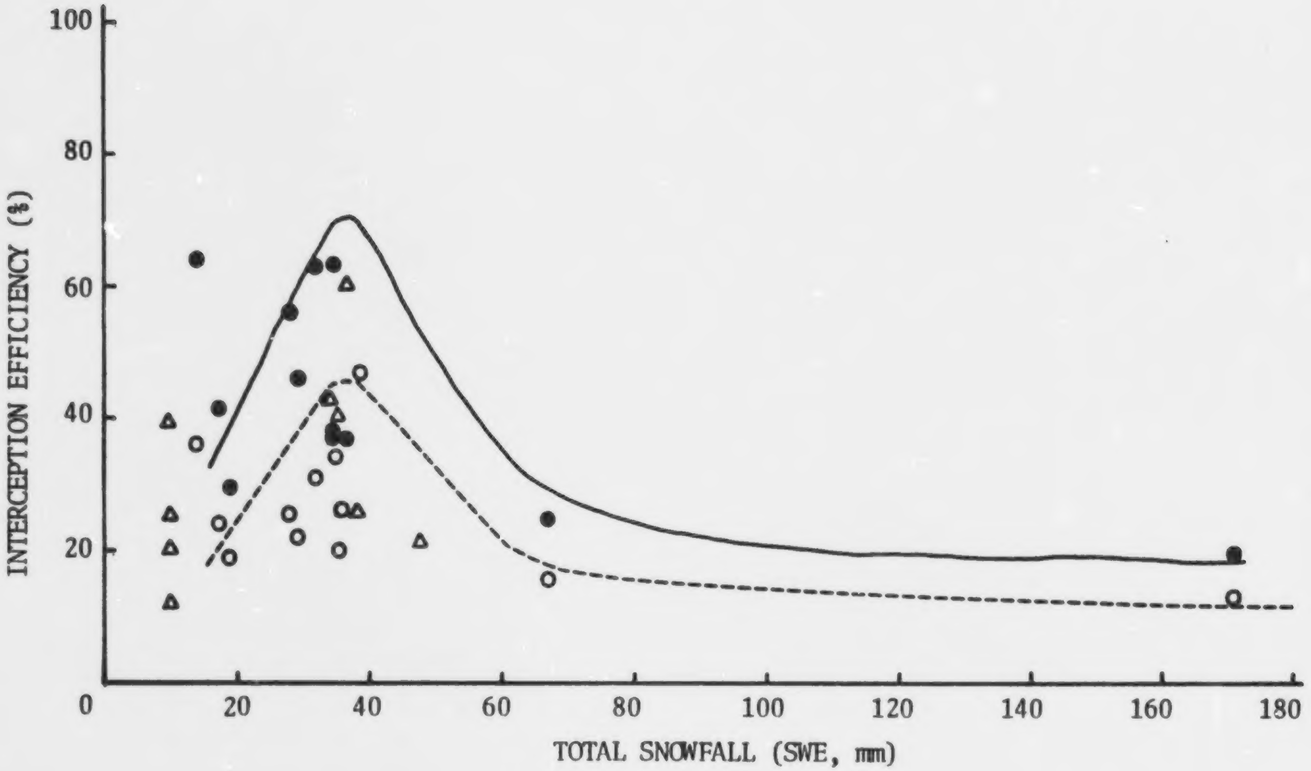


Figure 7.18 Interception efficiency as represented by the maximum snow load for storms of different sizes (● = Bokasugi, o = Kumasugi, Δ = sources in Table 7.2). (Bokasugi and Kumasugi data from Watanabe and Ozeki 1964: Table 3).

7.16). Different tree morphologies can alter dramatically the efficiency with which those trees intercept snow. The major influential features are the area and size of the component intercepting surfaces, angles of the surfaces, flexibility of the surfaces, inter-whorl distances and whole crown characteristics such as area, shape, and completeness. Different tree species and varieties exhibit different combinations of these features. We discuss each feature separately before summarizing the resulting interspecific differences.

7.6.1 Area and Shape of Component Surfaces

On a horizontal intercepting surface of infinitely large size, the amount of snow intercepted equals the snowfall. On smaller intercepting surfaces, however, the amount of snow intercepted becomes increasingly dependent upon the strength of cohesive forces which maintain a high angle of repose within the intercepted snow (e.g., Fig. 7.26a).

JGFES (1952) investigated the effect of surface size on interception by laying out horizontal boards of 0.5, 1, 2, 4, 8, 15, 30, 50, and 100 cm in width. The dotted line in Figure 7.19 shows the maximum amount of snow collected on different boards in a single storm. The widest board illustrated (30 cm wide) collected over 50-cm of snow whereas the 0.5-cm board collected only 4.5 cm. The solid line represents the amount of snow intercepted per cm of board width (Fig. 7.19). The

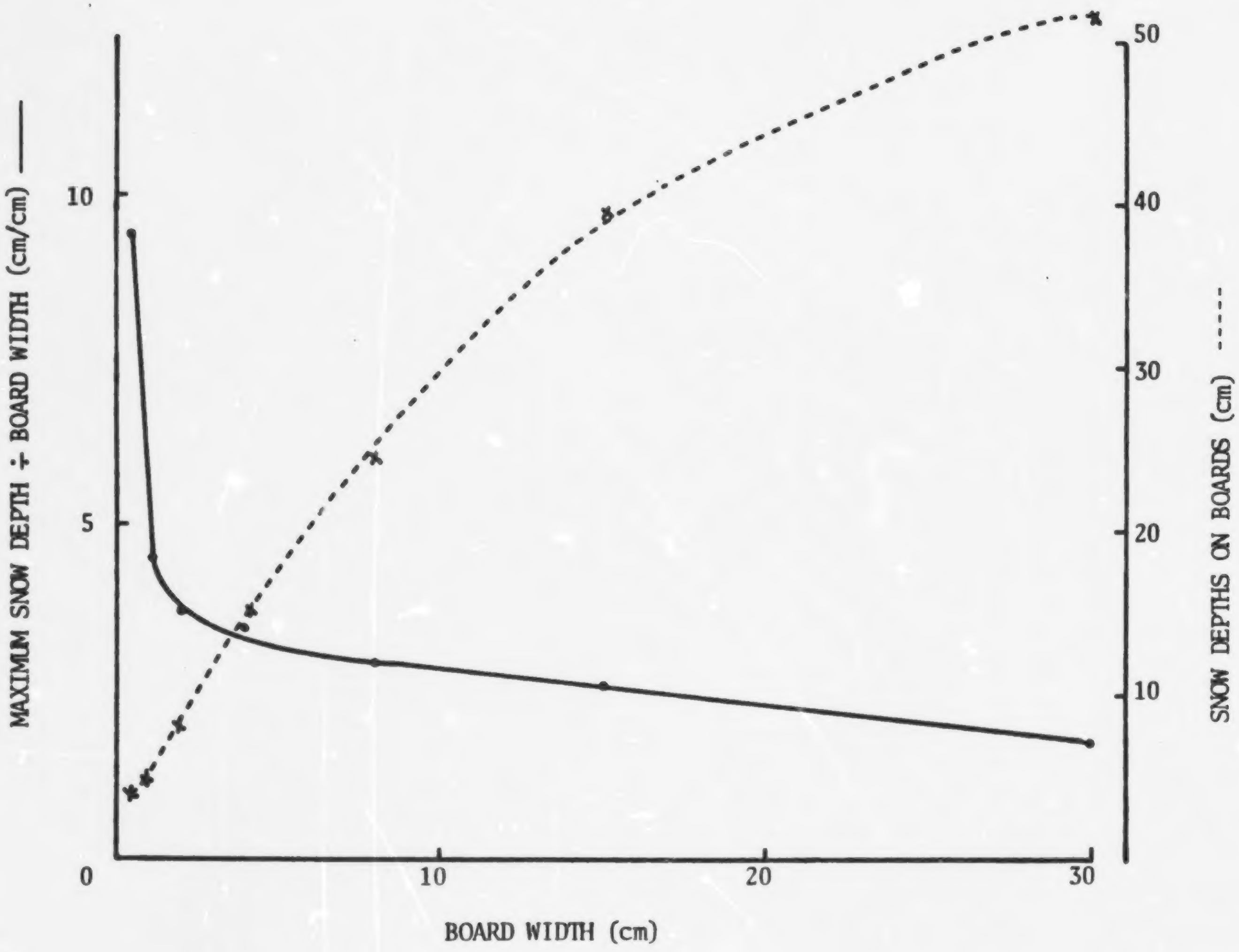


Figure 7.19 The influence of the width of flat, level boards on the maximum depth of snow caught and held (from data of JGFES 1952: 118).

relationship is strongly curvilinear with very small boards collecting far more snow than larger boards on a unit width basis. In further experiments a 16-cm board accumulated nearly the same depth of snow (96%) as the "standard", 100 cm wide board (Fig. 7.21).

The temporal pattern of accumulation of snow on flat boards during a storm (Fig. 7.20) differed from that within individual trees (Figs. 7.13-7.15). That is unsurprising because the effects caused by bridging of needles and crowns (Fig. 7.12) were not operative with the boards. The pattern illustrated (Fig. 7.20) represents accelerated metamorphosis of snow and more rapid cohesion with the passage of time. The dominant role of cohesion in generating the snow load has three important implications. First, the greater the snowfall, the greater the disparity between flat surfaces of different widths (Figs. 7.21a and b). Second, the elapsed time necessary for effective cohesion means that during moderate snowfalls narrower surfaces will retain disproportionately lesser amounts of snow than wider surfaces as length of the storm decreases (Fig. 7.21c). Third, the effect of temperature on cohesion and the resulting snow load will be more marked on narrower surfaces (e.g., Figs. 7.3 and 7.26).

Flat intercepting surfaces are more effective at intercepting snow than are cylindrical surfaces because the gravity vector is at right angles to the surface in the former. That in turn results in a small role for adhesion in

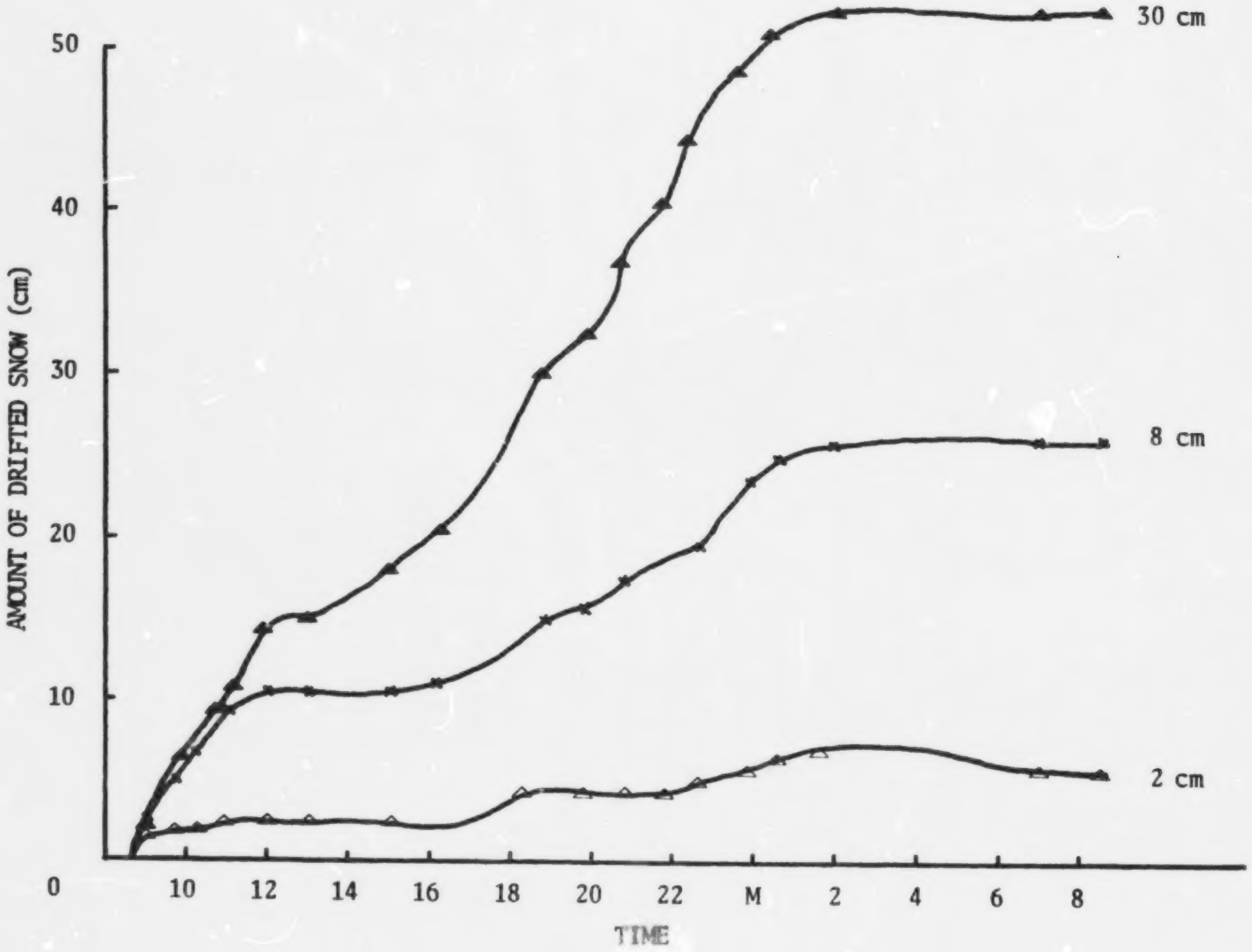


Figure 7.20 Temporal pattern of accumulated snow depth on flat, level boards of three widths during continuous snow storms (data from JGFES 1952: 118).

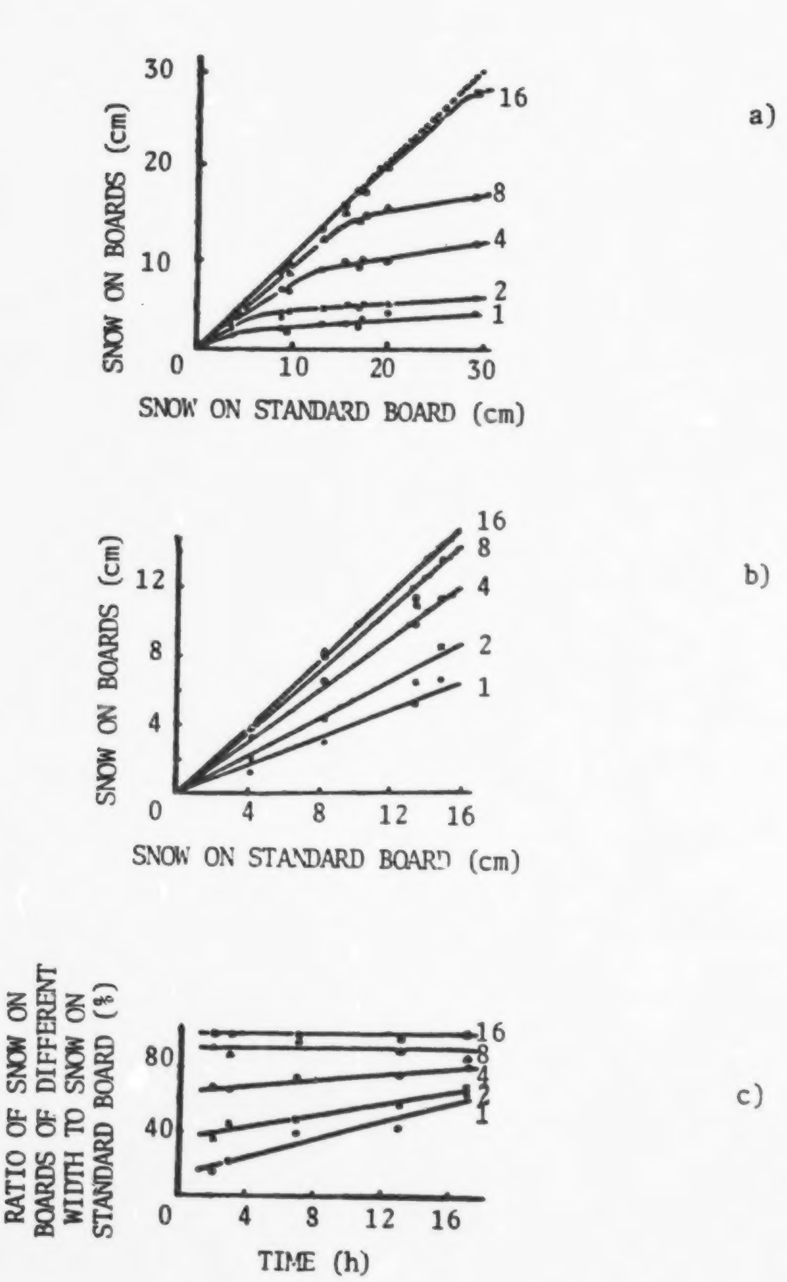


Figure 7.21 Temporal patterns of accumulated snow depths on flat level boards of different widths. Numbers refer to width of boards in cm (from JGFES 1952: 119).
a) Snow depths on boards of different widths versus that on a standard board (100 cm) during a heavy snowfall.
b) Snow depths on boards of different widths versus that on a standard board during a moderate snowfall.
c) Depth of snow on boards of different widths relative to that on a standard board as a function of time during a moderate storm.

the collection and retention of snow; cohesive forces dominate. In round branches snow slides off; the gravity vector is parallel to the intercepting surface at the outside edges. A rounded surface requires strong adhesive forces at all points, before snow can be retained effectively by cohesion. The JGFES investigated the effect of roundness by comparing maximal snow depths on round boards (i.e., dowels or poles) and flat boards where diameters of the round poles equalled the width of the flat boards. Very small dowels collected much less snow than small boards (Fig. 7.22). Larger cylinders, however, collected almost the same amount of snow as wide, flat boards. Over the range of diameters in data provided by JGFES (1952: 120) we calculate that depth of snow on a round board (Y) is related to depth on a flat board (X) by:

$$Y = 0.06\ X^{2\ 01} \qquad (r^2 = 0.98) \qquad (7.14)$$

Width and roundness tend to operate in opposition with respect to snow retention. Decrease in width of a rounded branch entails a relative increase in snow bearing capacity per unit width (provided cohesion is operating effectively), and a relative decrease in efficiency due to roundness. Figure 7.23a illustrates this trade-off and shows that round branches of about 2 cm in diameter held the greatest snow load per unit width. Data of Figure 7.23b suggest a smaller diameter for maximal accumulation per unit diameter, more

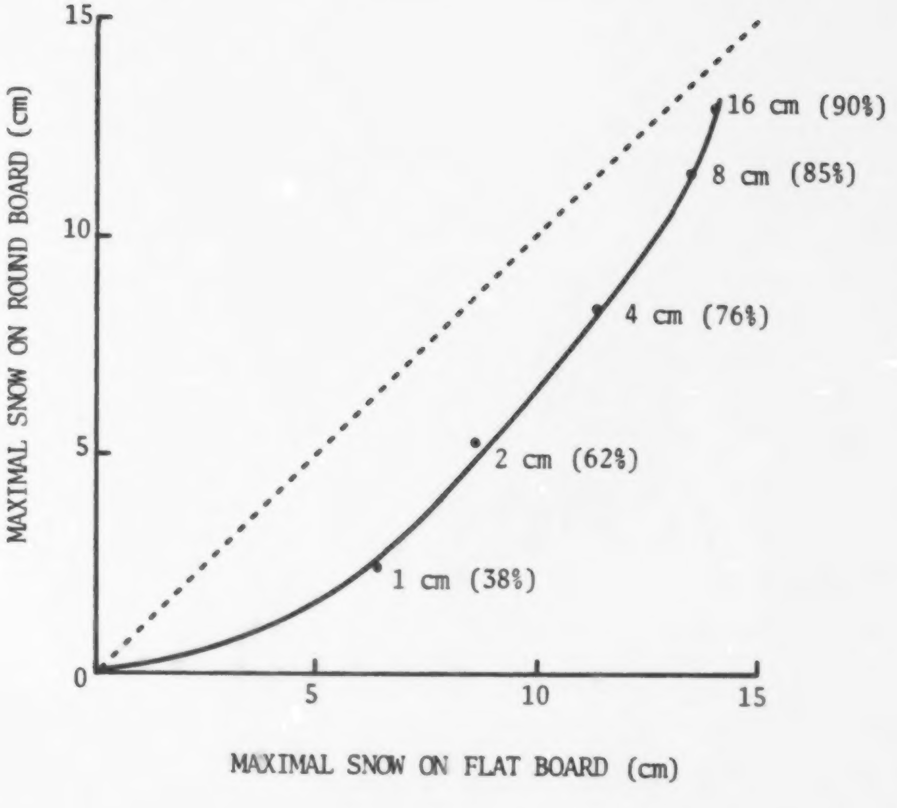


Figure 7.22 Comparison of the maximal snow depths accumulated on round dowels versus flat boards of 1.0 through 16.0 cm diameter or width (data from JGFES 1952: 120). Numbers in parentheses are the accumulations on a round surface as a per cent of accumulation of a flat surface.

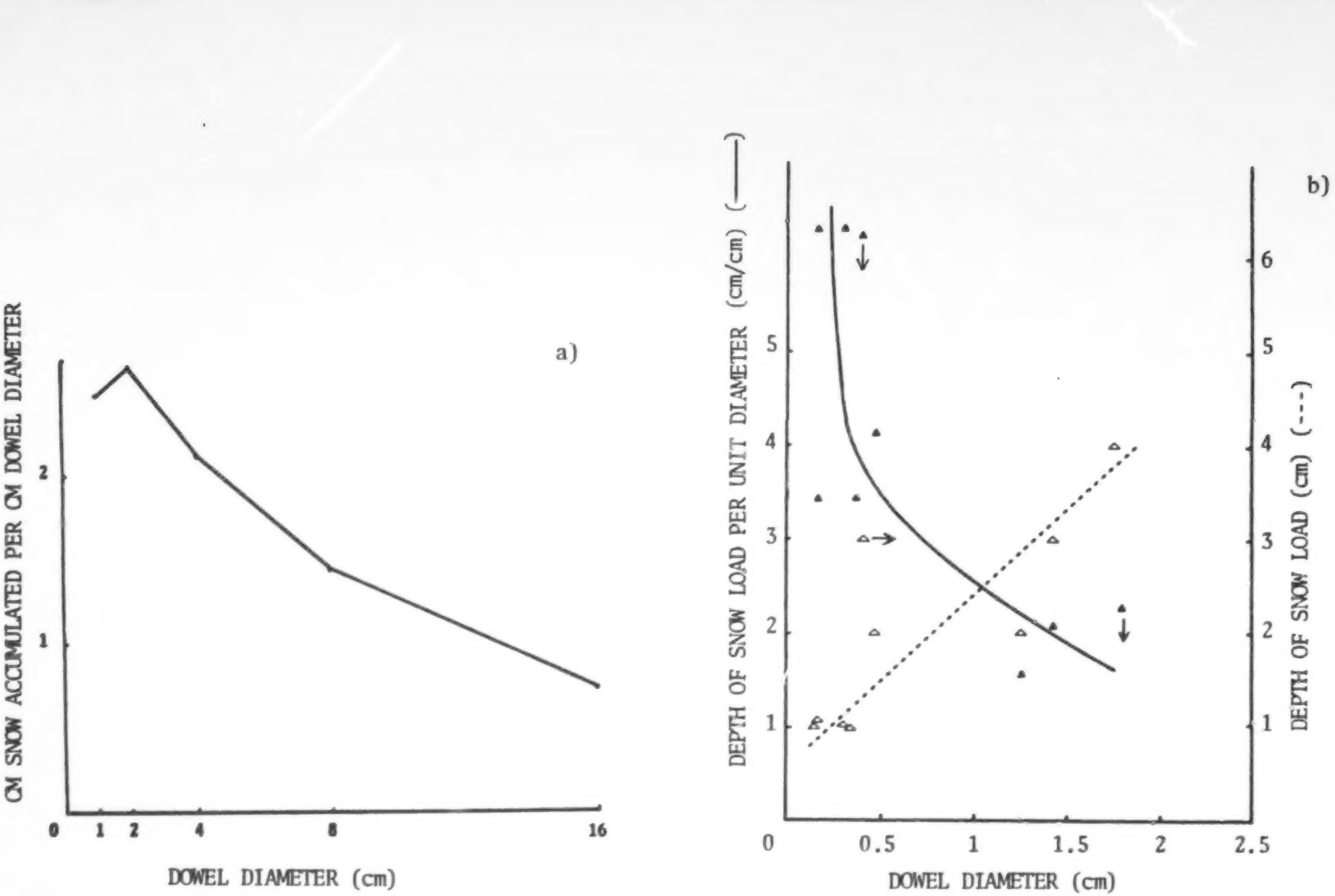


Figure 7.23 Snow accumulation as a function of diameter of rounded surfaces.
a) Maximum snow depth per cm of diameter of round dowels as a function of diameter (data from JGFES 1952: 120).
b) Maximum snow depth and depth per cm of diameter of round dowels as a function of diameter (data from Pruitt 1958: 170).

similar to that for flat boards (Fig. 7.19). These latter data are scaled from Pruitt (1958) and include values presented as more than some specific diameter (thus the arrows in Fig. 7.23b). Figure 7.23a is likely more accurate in form. It suggests that the optimum width for snow accumulation per unit diameter of rounded surface is larger than for flat boards on which adhesion is less important (Fig. 7.19).

These data suggest that for a given amount of intercepting surface on and in a tree crown, considerable snow would be intercepted by a large number of small intercepting surfaces even if those surfaces were rounded. A bare deciduous tree can be an efficient interceptor if it has an average branch width of about 2 cm (Fig. 7.23). A foliage covered tree, with many complex and flat surfaces, should hold a large snow load when the foliage is distributed over many, small surfaces; the load will still be less than that held by fewer larger surfaces. These conclusions do not incorporate the important effects of branch or leaf flexibility which may change the angle of interception (Ch. 7.6.2 and 7.6.3).

7.6.2 Angle of the Receptor Surface.

To investigate the effect of angle on snow interception by solid sloping surfaces, the JGFES (1952) laid out boards 90.9 cm in width at angles of 0°, 20°, 40°, and 60°. Maximum snow loads were measured on the boards as depth, weight per unit area, and density. We analyzed the data from 44 storms provided by JGFES (1952: 134). The effect of wind speed (V, $m \cdot s^{-1}$) on snow depth (H, cm) increased with increasing angle of the slope:

Angle	Relationship	
0°	$H = 16.08 - 5.25\ V$ ($r^2 = 0.08$, $SE = 6.75$, $P = 0.095$)	(7.15)
20°	$H = 16.77 - 5.60\ V$ ($r^2 = 0.09$, $SE = 7.01$, $P = 0.087$)	(7.16)
40°	$H = 19.44 - 8.04\ V$ ($r^2 = 0.14$, $SE = 7.92$, $P = 0.032$)	(7.17)
60°	$H = 21.22 - 12.12\ V$ ($r^2 = 0.21$, $SE = 9.06$, $P = 0.006$)	(7.18)

One corollary of the preceding relationships is that relative depth (H%, depth at a given slope relative to depth on the level) decreased dramatically with increasing slope and wind speed:

Angle	Relationship	
20°	$H\% = 105 - 17.8\ V$ ($r^2 = 0.02$, $SE = 45.3$, $P = 0.392$)	(7.19)
40°	$H\% = 123 - 21.9\ V$	(7.20)

$$(r^2 = 0.29,\ SE = 13.4,\ P = 0.001)$$

60° $$H\% = 134 - 59.7\ V \tag{7.21}$$

$$(r^2 = 0.62,\ SE = 18.4,\ P < 0.0001)$$

The steeper the angle, the greater is the effect of high wind speed in preventing the accumulation of snow depth (Figs. 7.24 and 7.28).

These experiments allow us to compute the rate at which the depth of accumulated snow decreases (DH%) with increasing angle of slope (θ, in degrees):

$$DH\% = 0.41\ \theta^{1.2} \qquad (r^2 = 0.98) \tag{7.22}$$

The equation holds best for moderate wind speeds of 1 to 2 $m \cdot s^{-1}$. Under those conditions a 10° angle would decrease the accumulated depth by 6.5%, a 30° angle by 24.3%, and a 60° angle by 55.8%.

One set of model trees utilized in JGFES experiments consisted of solid hexagonal pyramids. The different height:base ratios produced slopes of different angles. Figure 7.25 illustrates the effect of wind speed on snow depth in solid crowns of varying height to base ratios. Increasing wind speed caused a decrease in snow depth relative to that on a flat surface. The relationship is similar to that for the model of open cones with a small increment between levels (Fig. 7.5; small, inter-whorl distances of Eqs. 7.7 and 7.8).

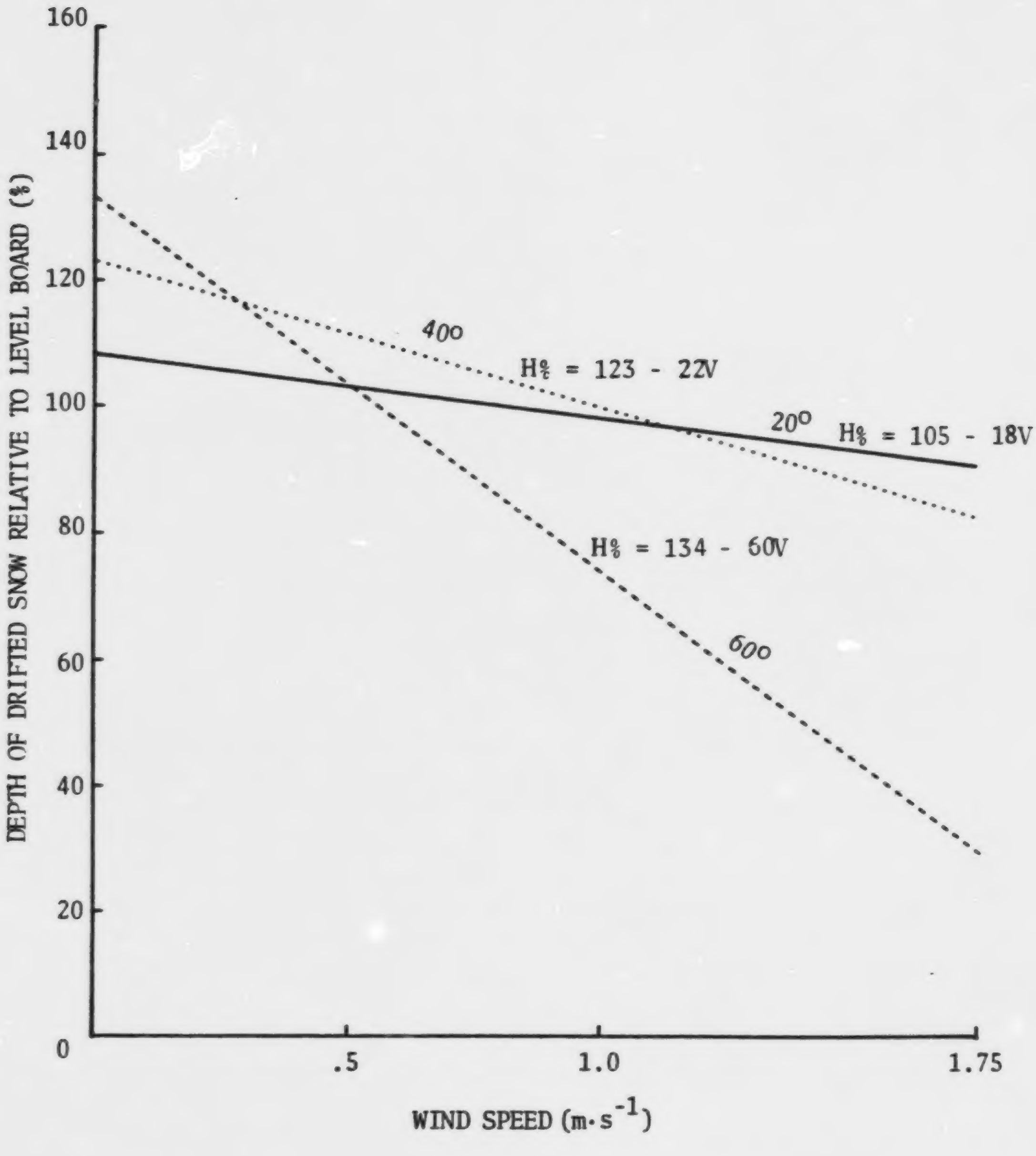


Figure 7.24 Effect of wind speed on the depth of drifted snow accumulated on boards of different slopes (from JGFES 1952: 136). H% is the relative depth of snow as a % of snow depth on a level board; V is wind velocity.

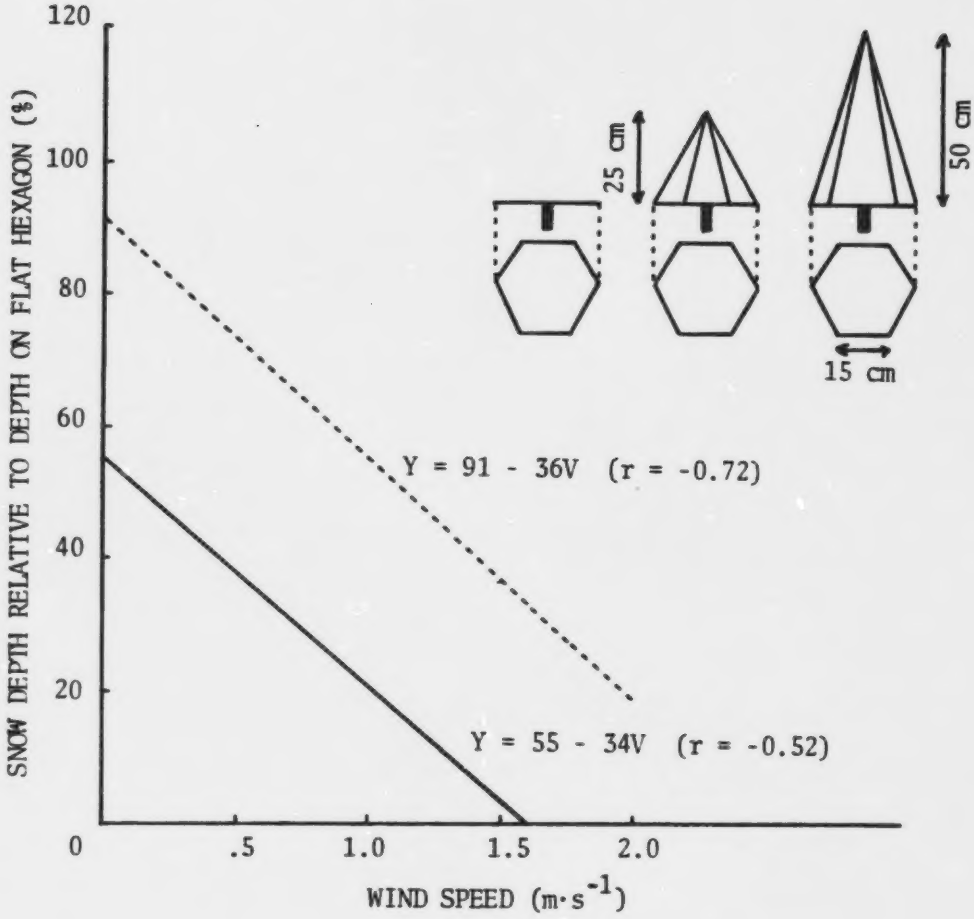


Figure 7.25 Effect of wind speed ($m \cdot s^{-1}$) on the accumulation of snow in modeled conical, closed crowns of different height:base ratios. Cone 25 cm in height, height to base ratio 1.67:1.00 (---); cone 50 cm in height, height to base ratio 3.33:1.00 (——) (data from JGFES 1952: 126).

Slopes of the relationship with wind velocity were the same but less snow was retained by the cone with the greater height:base ratio or more steeply sloping sides (Fig. 7.25).

During calm weather the experiment with the slanted boards produced the counter-intuitive result that intercepting surfaces at higher angles were characterized by greater depths of accumulated snow (Eqs. 7.15-7.21, Fig. 7.24). We expected the phenomenon to result from the fact that snow crystals on slopes bond with, and are supported by, other snow crystals not only on their upper and lower surfaces, but also on the uphill side (Fig. 7.26b). The result would be lesser densities and greater snow depths for the same SWE of intercepted snow on steeper slopes. Figure 7.27a illustrates that the explanation offered is consistent with the available data (JGFES 1952: p.134). At wind speeds $\leq 0.4\ \mathrm{m \cdot s^{-1}}$, in only 3 of 53 cases (5.7%) were densities greater on sloping surfaces than on the level (Fig. 7.27a). These findings appear to corroborate the conceptual model of Figure 4.1.

We can address predictions from that model more specifically by regressing either weight (W) or density (D) versus that predicted from the cosine of slope angle (Θ) under conditions of low wind speed ($V \leq 0.25\ \mathrm{m \cdot s^{-1}}$). The model (Fig. 4.1) is based on vertical snowfall and that is unlikely to be approximated even in calm air. Observed mean densities (Do) declined linearly with slope angle for the 11 storms when wind speed was $< 0.25\ \mathrm{m \cdot s^{-1}}$:

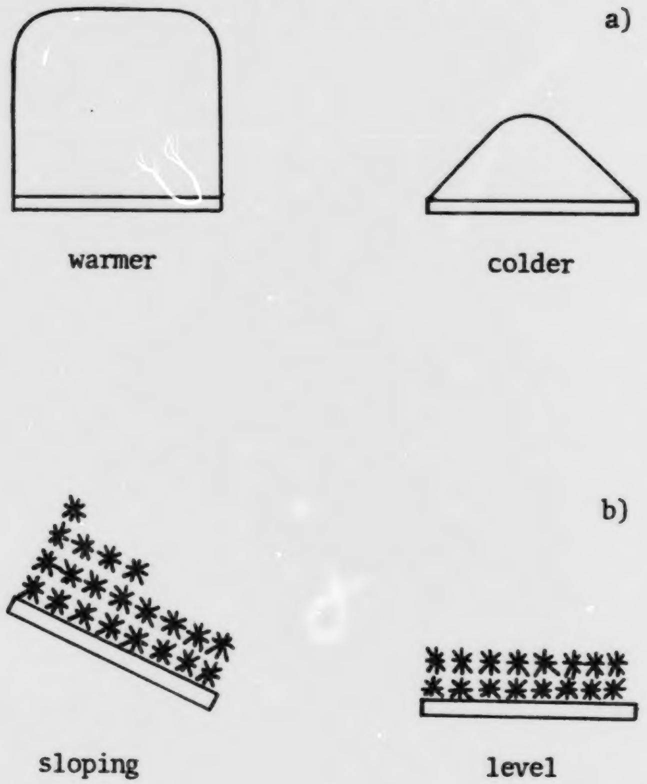

a)
warmer
colder
b)
sloping
level

Figure 7.26 Temperature and slope effects on cohesion.
a) On level surfaces temperatures just below zero will produce the greatest snow depths.
b) Provided wind speed is low, sloping surfaces should produce greater snow depths than level surfaces. Wind is from the right.

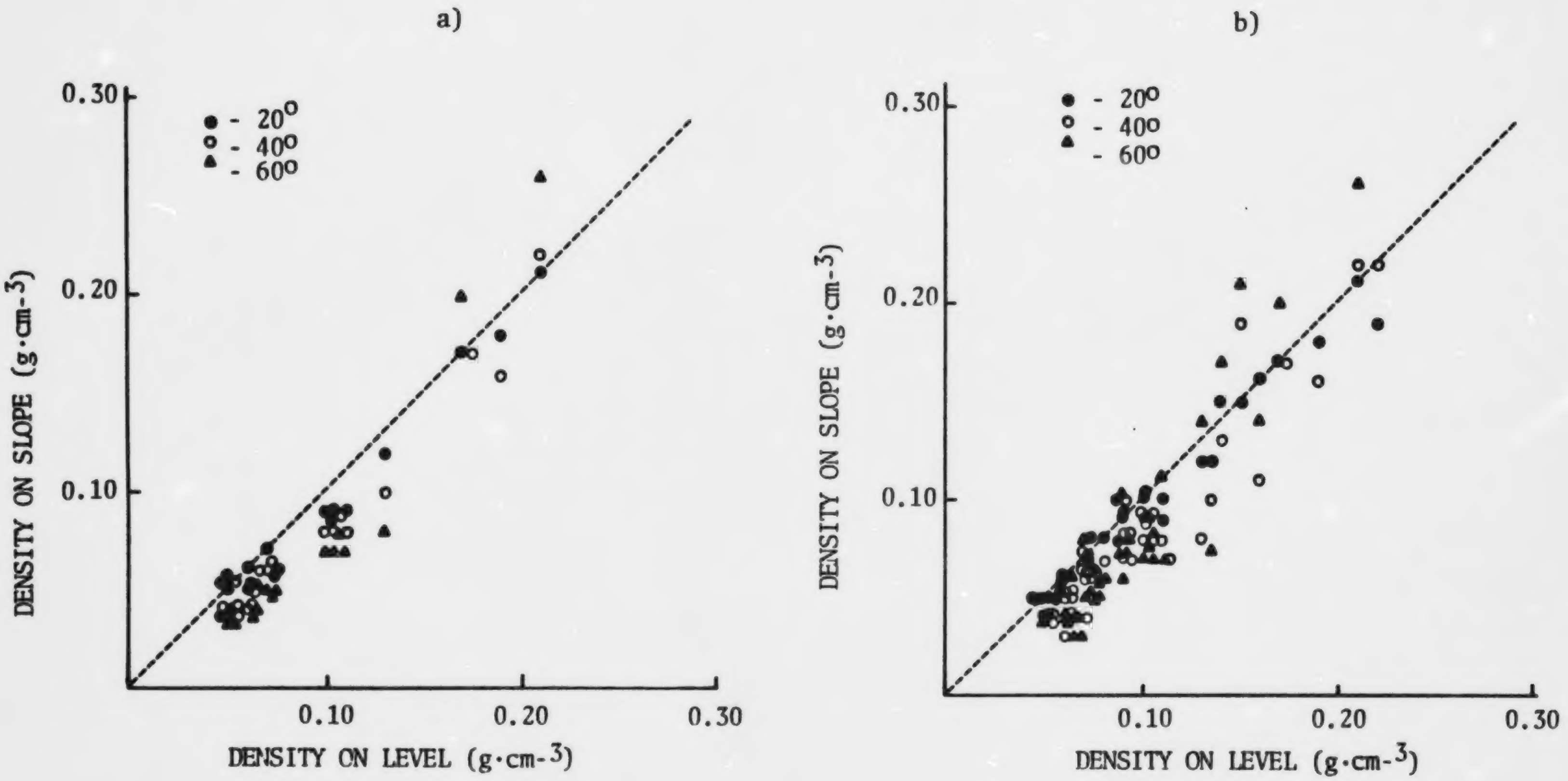


Figure 7.27 Densities of snow on sloping boards versus densities on a level plane (data of JGFES 1952: 134).
a) Wind speeds less than 0.4 m·s-1.
b) All wind speeds combined.

$$Do = 0.073 - 0.00035\ \theta \qquad (7.23)$$

$(r^2 = 0.996,\ SE = 0.0007,\ P < 0.002)$

Given the apparent linearity, mean densities exceed those predicted by cosine of the slope angle (Dp):

$$Do = 0.03 + 0.53\ Dp \qquad (7.24)$$

$(r^2 = 0.90,\ SE = 0.0035,\ P < 0.05)$

Because of the opposing effects of snow depth and snow density, weight of snow showed no relationship with slope angle ($r^2 = 0.16$, $P > 0.5$). Density follows the direction but not the relationship predicted, weight does not. The model of Figure 4.1 predicted only snow delivery and did not consider adhesive and cohesive forces. Cohesion is probably more effective in encouraging snow accumulation on moderate slopes (Fig. 7.26b), such that gravity does not play a dominant role until slope angle exceeds about 65° (e.g., Fig. 7.29).

Further predictions regarding the interaction of wind speed and slope can be extracted from the model of Chapter 4 (Figs. 4.1 and 4.2). The model related to large, topographic or cloud-forming surfaces and the predictions must be modified somewhat for the smaller surfaces, such as those comprising a tree crown, especially because these latter surfaces can also be shaken by wind. The predictions considered are:

1) Snow depths on windward slopes relative to depth on the

level will decrease with increasing angle of the slope (Fig. 4.1) and increasing wind speed. High wind speeds will overcome the positive slope effect of Figure 7.26b and the effect of wind speed will be more pronounced on steeper slopes. As intensity of the snowfall increases, the effect of wind speed should be reduced, particularly on shallow slopes (e.g., Fig. 4.2).

2) The effects of wind speed on snow density will change with angle of the slope. In snow storms of low intensity less steep slopes should show less change in density with increasing wind speed. As the slope steepens to become more perpendicular to the wind, the density should increase markedly with increasing wind speed (Fig. 4.2).

3) At storms of high intensity the potential influences of both slope and wind speed on snow density should be reduced. Accumulated weight of the snow itself will lessen the potential impacts of wind speed on density.

4) On sloping surfaces experiencing moderate wind speeds, the weight of snow per unit area should be greater than on the level when rates of snow production are low and smaller when rates of snow production are high (because of the diagonal vector of snowfall, Fig. 4.2)

Variables available for evaluating these predictions are incomplete surrogates. However, both parts of the first prediction are met. Depth decreases faster on greater slopes with increasing wind speed (Fig. 7.24); at storms of greater intensity the effect of wind speed is reduced more on shallow slopes than on steeper slopes (cf. Fig. 7.28a and b).

The response of snow density to wind speed is to some extent the opposite of that of snow depth. Density appears to increase more on steeper slopes with increasing wind speed (Fig. 7.27b); that finding is likely a function of greater impact velocity and compaction. When densities of snow on the level were less than 0.13 $g \cdot cm^{-3}$ (usually near-calm conditions, cf. Figs. 7.27a and b), densities of snow on the slanted boards were consistently lower, particularly for steeper slopes. We evaluated data for 44 storms and found no statistically significant influence of wind speed on density for any specific slope. Because the effect was greater on steeper slopes, the relative density (D%) did increase significantly with wind speed on the steepest slope (60°):

$$D\% = 0.735 + 0.24\ V \qquad (7.25)$$

$$(r^2 = 0.20,\ SE = 0.186,\ P = 0.0075)$$

When storms were segregated by intensity, the second and third predictions were also met. The interaction of the effects of wind speed and slope on snow density was evident primarily in storms producing 1 to 10 cm SWE (cf. Fig. 7.28c and d). When all storms and slopes were combined ($n = 132$)

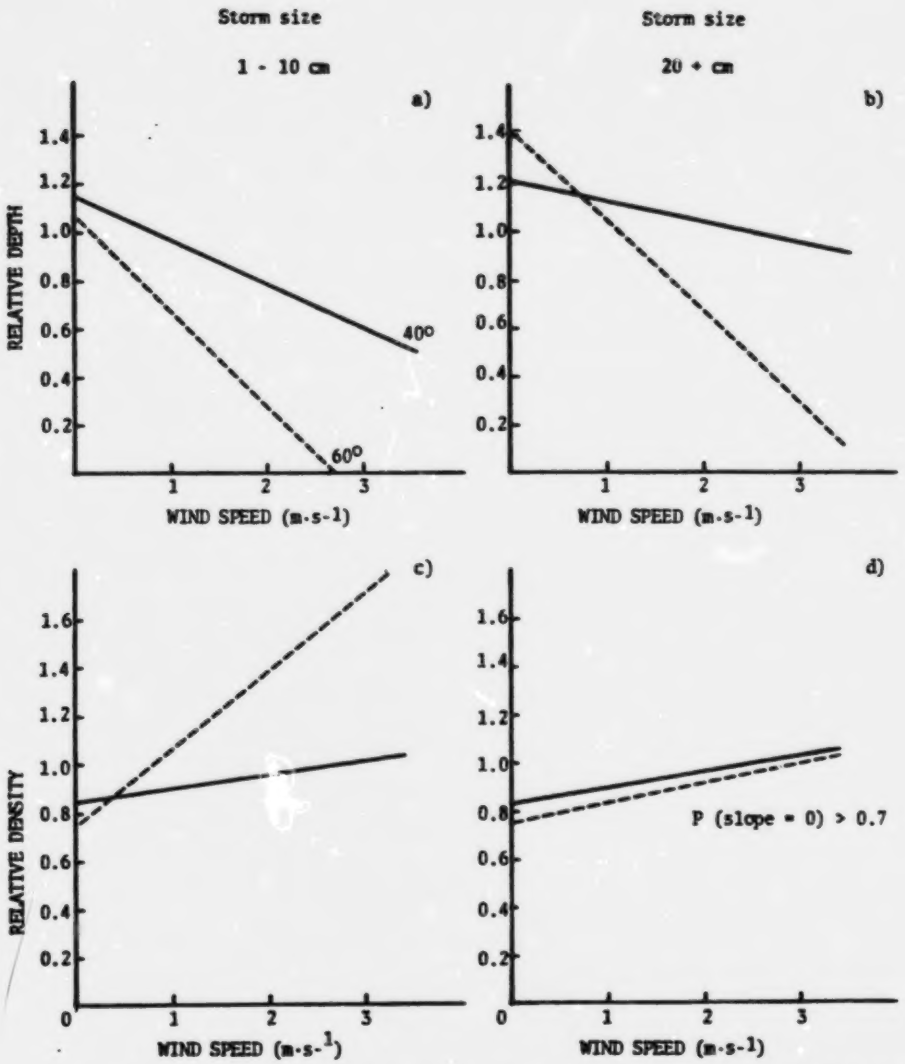


Figure 7.28 Interaction of slope angle and storm size on the effects of wind speed on snow depth and density. Relative measurements are relative to those on a level surface (regression from data of JGFES 1952: 134).

a) Influence of wind speed on snow depth in small snow storms.
b) Influence of wind speed on snow depth in large snow storms.
c) Influence of wind speed on snow density in small snow storms.
d) Influence of wind speed on snow density in large snow storms.

there was no relationship between wind speed and snow density ($r^2 = 0.03$).

The predictions regarding weight were not met, largely because of the opposing effects of wind speed on depth and density. Slopes of regressions of relative weight versus wind speed were negative for all combinations of slope and storm size evaluated, but in no case were they significant ($P > 0.05$).

The effect of angle has opposing effects on snow depth and snow density that are further modified by wind speed and storm size. In calm weather the depth of snow is greater on steeper slopes (Fig. 7.24) but the density is generally less (Fig. 7.27a). As a result, the weight of snow on a unit area is largely unrelated to slope of the area over a large range of angles. Summarizing data from the solid pyramids (Fig. 7.25) and slanting boards (Fig. 7.24) yields the following observations in terms of weight of snow:

	Angle of surface				
	20°	40°	60°	63°	78°
ratio of snow weight accumulated to that of a flat surface (%)	98	96	89	91	55

These data are illustrated in Figure 7.29. They suggest a rapid decrease in efficiency of interception at angles of about 65° or a crown height to base ratio of about 2.15 to 1.

The relationship of snow load (kg) to angle is more complicated in windy weather (Fig. 7.28). However, the general pattern is that increasing wind speed tends to reduce

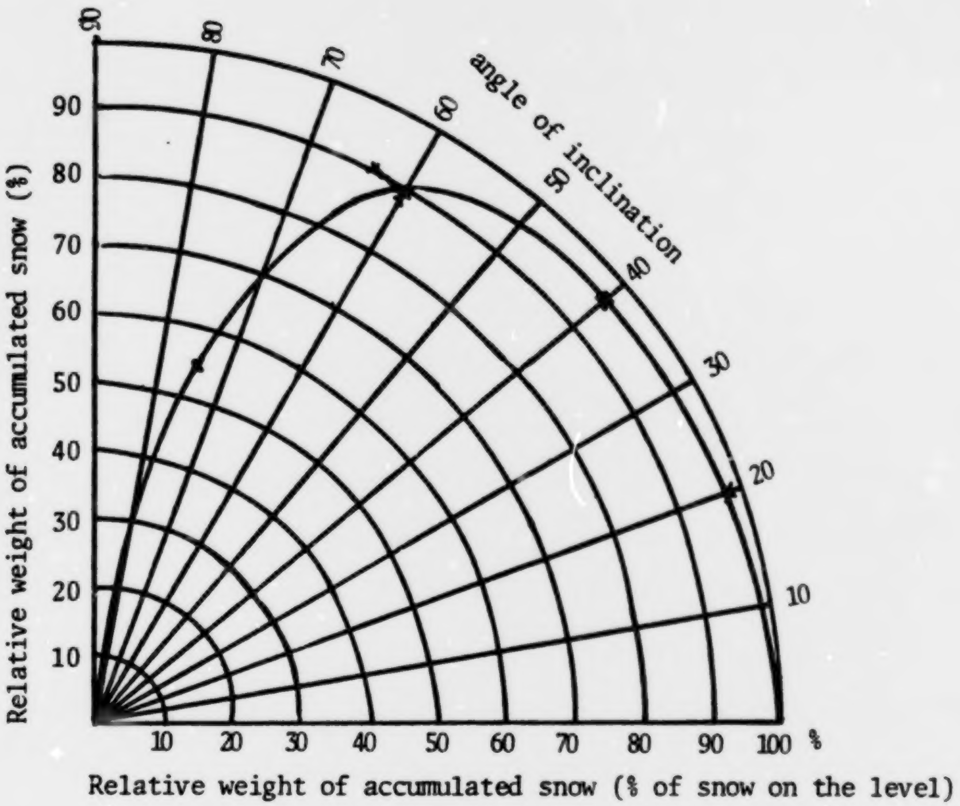


Figure 7.29 The weight of snow accumulated on slopes of varying angles relative to weight of snow on a level surface during calm weather (data of JGFES 1952: 128).

snow load on trees as the angle of the intercepting surface increases (Fig. 7.24). As a result, in pyramidal or conical crowns, measured snow load decreases with wind speed (Fig. 7.7). Note that the effects of wind speed in reducing snow will be greater in trees with flexible branches than in model crowns of solid surfaces.

The relationships of Figures 7.7, 7.24, 7.25, and 7.29 suggest one reason why young plantations are less effective at intercepting snow than are old-growth forests: the large height to base ratio of young, densely stocked trees provides an inefficient surface for interception simply in terms of angle. Furthermore, the generally shorter, mean inter-whorl distances of young trees (particularly those that have undergone little self-pruning or experienced no breakage) approximate models 2 and 3 of Figure 7.5. The relationships also suggest that the extremely large height to base ratios found in tree species of the boreal and sub-alpine forests may be an adaptation encouraging snow shedding (not protection from ultra-violet radiation as suggested for subalpine species by Zinke 1968 (pers. commun. in Anderson 1970). Ignoring other effects, silvicultural practices which increase the crown height:base ratio to produce steeply sloping sides may reduce snow interception. At the broadest level, management for 'piece-size' as opposed to 'volume' may increase snow interception despite the wider spacing.

The JGFES (1952) suggested that the general relationship between snow depth, angle, and wind speed could be expressed as:

$$S(\theta) = S(o)\ (a + b\ \mathrm{Sin}\theta + c\theta + V) \qquad (7.26a)$$

or:

$$S(\theta)/S(o) = a + b\ \mathrm{Sin}\theta + c\theta + V \qquad (7.26b)$$

where θ = angle of intercepting surface, $S(\theta)$ = snow depth at angle θ, $S(o)$ = depth on a horizontal plane, V = wind speed and a, b, c, and d are constants depending upon snow and intercepting surface characteristics. For their data: $a = 97$, $b = 41$, $c = 0.41$, $d = 1.2$.

7.6.3 Flexibility of the Intercepting Surface

As snow load increases on a tree's intercepting surface, the leaves and branches bend to assume a more vertical orientation. This change affects potential interception in two ways. First, the effective surface area of the tree is reduced as the crown is deformed into a cone with a larger height:width ratio (e.g., Fig. 7.29). Second, as the angle increases, the gravity vector becomes more effective in overcoming adhesive and cohesive forces; less snow is accumulated and more snow is dropped (Ch. 7.6.2). The flexibility of a surface or its resistance to bending thus determines its effectiveness in supporting adhering snow.

Apart from the height:width ratios noted earlier, increased flexibility of leaves and branches is the primary manner in which natural selection has acted to adapt tree species or varieties to conditions of heavy snowfall. Anecdotal examples abound in the literature of regional phenotypes varying in their branch flexibility as a response to snow conditions (see Ch. 7.6.5).

However, little experimental or quantitatively descriptive work has been done on the effect of leaf or branch flexibility on snow interception. Lull and Rushmore (1961) used time-lapse photography to record that white pine needles accumulated snow only at the base of the fascicles before the needles were bent into a platform. Of the four species Lull and Rushmore studied, balsam fir, with many stiff branchlets and needles persisting along the branches, was the best collector of snow; hemlock was the poorest collector because its needles were "feathery and flexible". The JGFES made an interesting comparison of maximal snow and sawdust loads on a naturally flexible cedar branch versus on a rigid, complex, slatted, foliage model (about 45 by 25 cm; JGFES 1952: 150). The rigid model had a much higher maximal snow load than the cedar branch (simulated by sawdust in Fig. 7.1).

The only workers who have quantified the effects of branch flexibility on interception appear to be Watanabe and Ozeki (1964). They studied two varieties of Cryptomeria japonica: Bokasugi from Toyama prefecture and Kumasugi, a snow-adapted variety from Niigata prefecture. The Kumasugi was 23 years old with a crown length of 6.1 m. Vertical projected area of the crown was 8.6 m^2; horizontal projected area, 5.0 m^2. The total length of primary branches was 91.5 m, and the average angle of their attachment was 76°. Two Bokasugi trees were used in the experiments. Their measurements were: age (35 years), crown length (2.4, 2.3 m), vertical and horizontal projected areas of the crown (4.0, 3.8 and 2.4, 2.3 m^2, respectively), total length of primary branches (54.6, 51.3 m), average angle of primary branch attachment (53, 55°). On the basis of measurements of crown area and length, and length of branches we would expect the Kumasugi to catch and retain more snow. In fact, the Kumasugi accumulated the smaller snow load primarily as a result of the angle of its branches and their flexibility.

As crown snow load during a storm increased, the less flexible branches of the Bokasugi were depressed only slightly from about 54 to -80° where 90° equals horizontal (Fig. 7.30a). The more flexible branches of Kumasugi, which had an initially more favourable angle, were quickly depressed to -44° under the same snow load (Fig. 7.30a). The change in angle of the foliage reduced the vertical and horizontal projected crown areas of the Kumasugi to about 70% of their initial value (Fig. 7.30b). Because the Bokasugi branches were depressed only slightly below a horizontal plane, the projected crown areas were actually increased to about 120% of their initial values. The Bokasugi approximated the open conical models of Figure 7.5, whereas the Kumasugi approximated the pyramidal models (Fig. 7.25). As a result the Bokasugi accumulated a much greater snow load despite their smaller crown area.

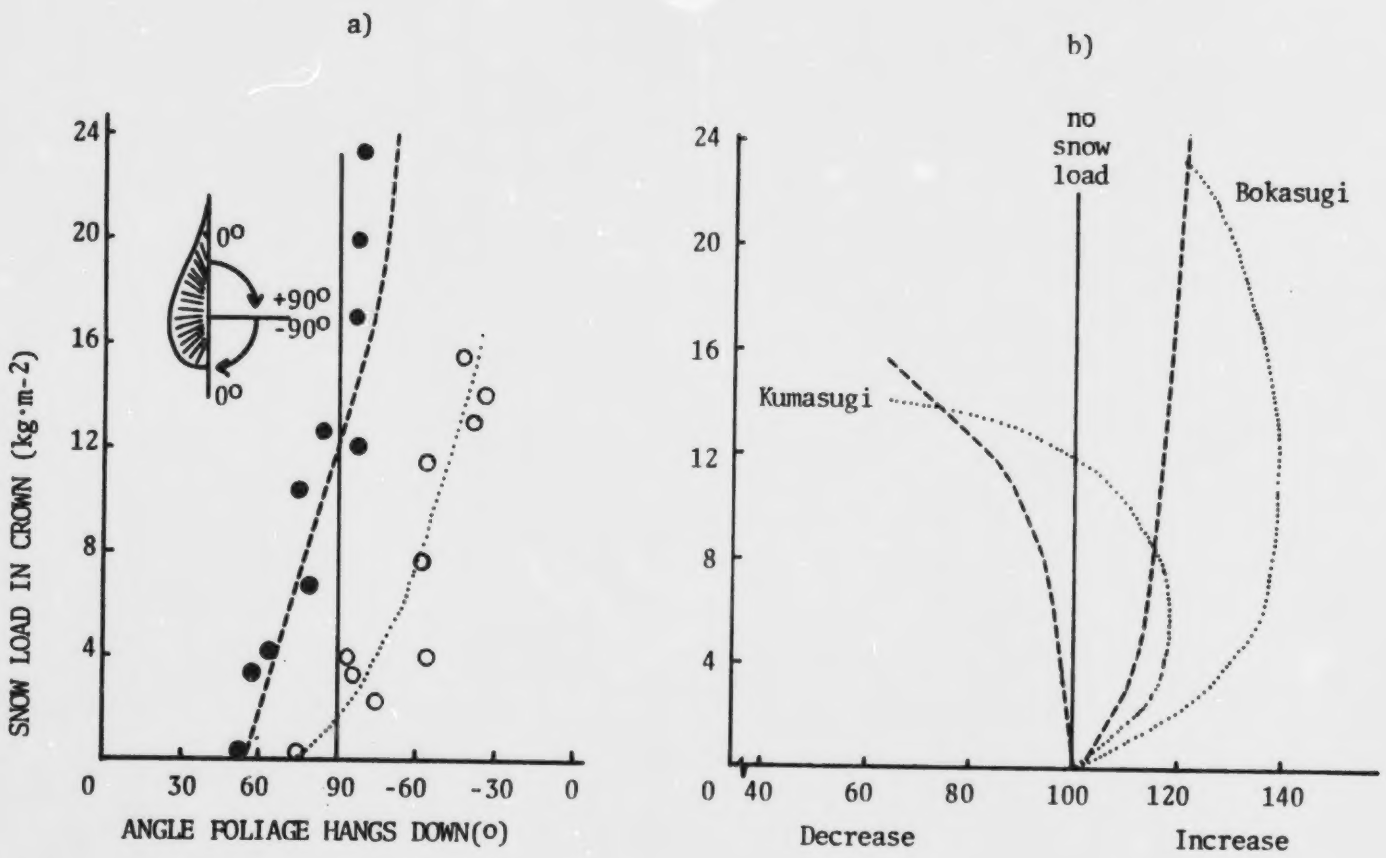


Figure 7.30 Change in the shape and area of crowns of two *Cryptomeria* varieties as a function of snow load (data of Watanabe and Ozeki 1964: 126).

a) Relationship between snow load and the angle at which foliage hangs down; ● = Bokasugi, o = Kumasugi.

b) Change in the projected area (shape) of crowns as a function of snow load; ----- vertical projected area, ····· horizontal projected area.

The pattern of snow accumulation on these two varieties was illustrated earlier for the same storm presented in Figure 7.30 (Fig. 7.15). In that storm the Bokasugi accumulated 23.3 $kg \cdot m^{-2}$; the Kumasugi, 14.1 $kg \cdot m^{-2}$. We calculated the mean interception efficiency during 12 storms to be 47.4% for Bokasugi and 16.9% for Kumasugi (Watanabe and Ozeki 1964: Table 3). In further studies on the potential effects of tree age on interception efficiency, Watanabe and Ozeki documented the generality of the effect branch flexibility has on crown area and the consequent reduction in snow load (Fig. 7.33).

The three studies cited quantify the importance of surface flexibility to the interception of snow (JGFES 1952, Lull and Rushmore 1961, Watanabe and Ozeki 1964). In each case rigid surfaces retained a greater snow load than flexible surfaces. In the case of the two Cryptomeria varieties the effect of increased flexibility was not only to reduce the projected area of the crown (Fig. 7.30b), but also to depress the branches and introduce negative slope (Fig. 7.30a). The effects of slope on snow interception (Ch. 7.6.2) were then initiated. Although the results indicate an important effect, it is difficult to refer them to other species. There are two clear implications to methodology in studies of snow interception. First, crown measurements taken in the absence of snow load may be markedly different from the shape of the crown when carrying snow. Second, the degree to which crowns alter their surfaces differs between species and with age in the same variety (Fig. 7.33). Thus crown measurements taken in the absence of snow may misrepresent real differences that influence interception efficiencies.

7.6.4 Whole Crown Attributes

The effects of size, shape, angle, and flexibility of the component surfaces of a crown obviously have important influences on the efficiency with which individual crowns intercept snow and the maximum snow load they can hold (Ch. 7.6.1-7.6.3). For management purposes simpler, integrative measurements are clearly preferable. Many have been proposed; we have grouped them broadly as linear, areal, and volumetric. To relate any specific measurement to interception by a single tree one must first measure interception by the tree. Standard comparisons of snow in the 'open' versus 'under the canopy' are inadequate for single trees because the angle of snowfall determines snow depth under the tree and redistribution of intercepted snow can occur (Ch. 6). Only continuously weighed trees are appropriate to the question. Such data are very sparse and we have relied heavily on reanalysis of data provided by Watanabe and Ozeki (1964). Our approach is first to review candidate measures, then present experiments and results of Watanabe and Ozeki, and finally to discuss each candidate measurement.

Candidate measurements.--Among the linear measurements which have been proposed are age (because it can be quantified), crown length (length of live crown), number of branches, total length of branches, and the angle of branch attachment. The first two can be measured relatively easily. They are the only two that can be considered strictly linear; that is, representing only one dimension. Total number of branches and the total length of branches are more representative of potential intercepting area. Angle of branch attachment also represents area as it determines how the branches are presented relative to snowfall. The definition of angle of attachment of primary branches follows that of Figure 7.30. Areal measurements ideally would encompass all the intercepting surfaces of the crown. Because that is impossible, several surrogates that reflect crown area or shape have been proposed. These include the angle, number, and length of branches as noted, crown height to base ratio, vertical projected area, horizontal projected area, and outside surface area of the crown. Height to base ratio is more clearly a measure of crown shape and reflects the relative proportions of vertical to horizontal area. Horizontal projected area (HPA) is simply the area on the ground (in a horizontal plane) which would be covered by the crown if one were looking straight down. It incorporates no measurement of crown completeness. In our calculations of height:base ratios we have assumed that trees were circular at the base of the crown; we then could compute crown diameter from HPA. It is clear from the data of Satterlund and Haupt (1967) that they derived HPA from crown diameter assuming a circle. We cannot evaluate whether Watanabe and Ozeki (1964) determined HPA in the same fashion.

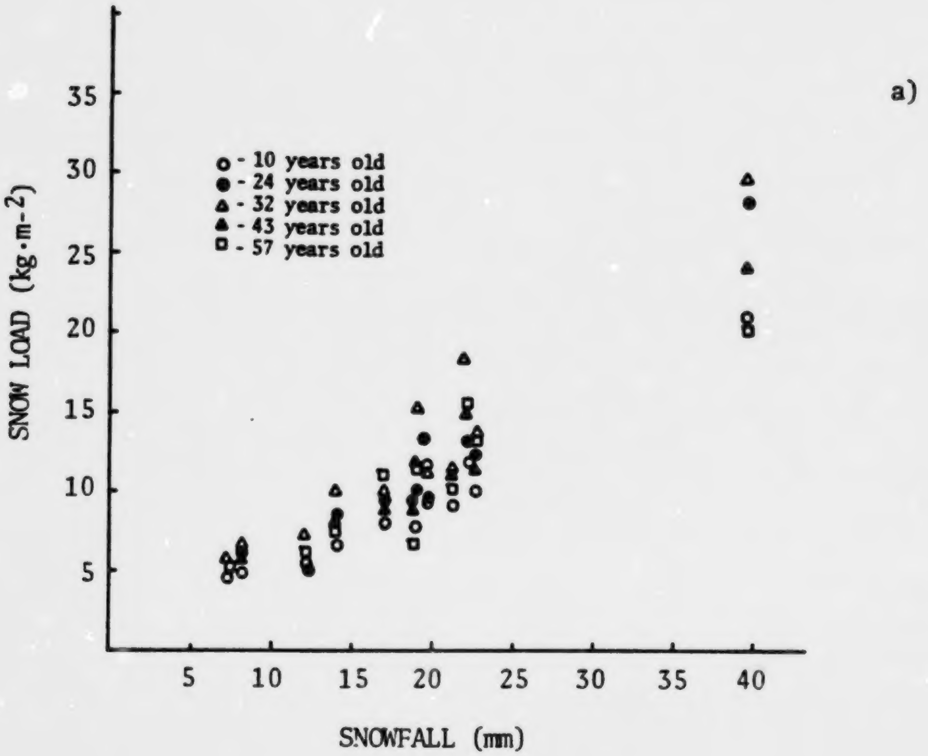
Vertical projected area (VPA) represents the area projected by the crown in plan view. Our calculations assume it represents a triangular section through a cone. To calculate outside area of the crown, termed crown surface, we have assumed crowns were conical and derived the base of the cone (crown diameter) by assuming that HPA was circular. The height of the cone was simply crown length. We derived and evaluated three indices of the density of the surface. These were the total primary branch length divided by HPA, VPA, and crown surface. The three indices were assumed to represent an index of canopy completeness.

One might assume that volumetric measurements relate best to processes of interception because they more accurately reflect all receptor surfaces. However, they too are impractical. We evaluated the potential utility of two indices. The first, crown volume, estimates the volume assuming a conical shape as was done to estimate crown area. The second represents an approximation of canopy completeness. We term it crown density and it is derived by dividing the total length of primary branches by the estimated crown volume. It is the primary branch in meters length per cubic meter of crown.

The three measures of interception used are total snow load (kg, as reported), snow load per m^2 ($kg \cdot m^{-2}$; as reported, where area is the HPA), and interception efficiency (%) which we derived as a function of measured snowfall assuming a density of 0.10 g·cm^{-3}. The mean (±SD) of densities for the 26 storms reported was 0.096 ± 0.026 g·cm^{-3}, so on average our calculations of interception efficiency underestimate by 4%.

Experimental data.--The two experiments from which most of the data were derived were pursued from 1957 to 1963 by Watanabe and Ozeki (1964). The reanalyses presented here represent our extensions of their findings. Where possible in our summary treatment we have included appropriate data from other continuously weighed trees (JGFES 1952, Satterlund and Haupt 1967).

In one experiment five Kumasugi (snow adapted) Cryptomeria trees of different ages were weighed during 12 storms. The trees' ages were 10, 24, 32, 43, and 57 years; all had a crown length of 4.8 m. Photos indicate they all had the same height and were positioned, in the weighing device, the same distance above ground. Over the 12 storms there were few clear patterns as a function of age (Fig. 7.31). In fact, the 32-year-old tree tended to show higher snow loads per m^2 and greater interception efficiency. Figure 7.31a does illustrate the consistent exponential buildup of snow load with increasing but still moderate snowfalls. That is probably largely a function of bridging between intercepting surfaces (see also Fig. 7.12 and 7.17). The interception efficiency (Fig. 7.31b) shows consistently high values with wide scatter. The range of snowfall (storm sizes) illustrated approximates



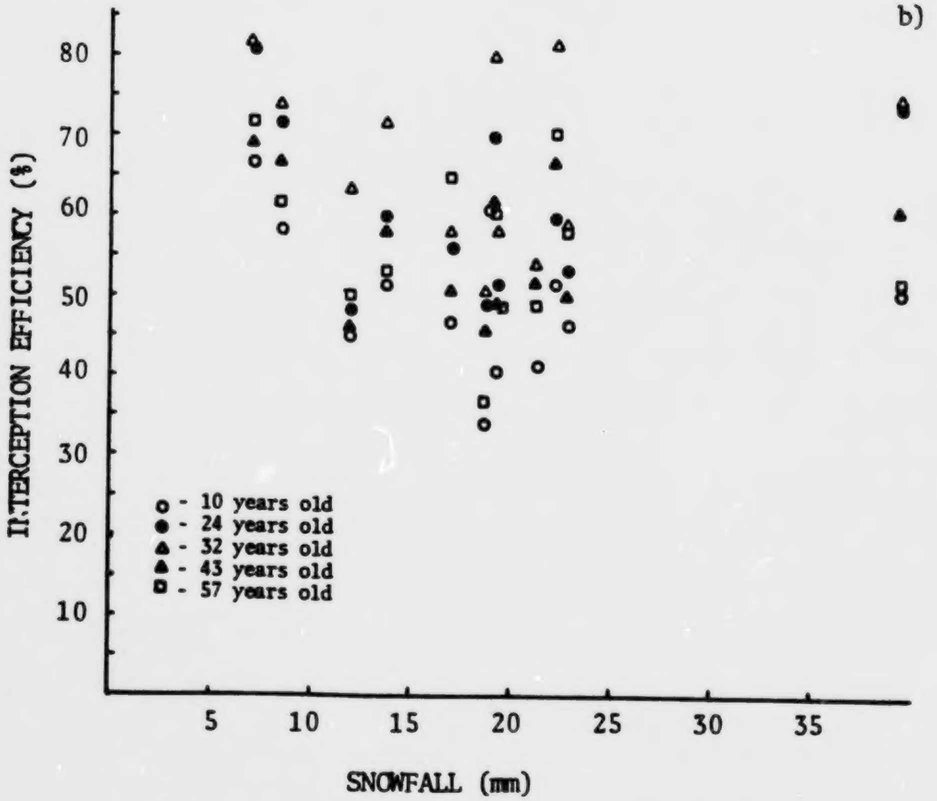


Figure 7.31 Interception by Cryptomeria japonica Kumasugi trees as a function of age and snowfall (mm). (data from Watanabe and Ozeki 1964: 128, 129).
a) Snow load (kg·m-2).
b) Interception efficiency (%).

the range of high interception efficiencies illustrated earlier (Fig. 7.18).

To examine which crown attributes might most influence interception we compared average values for 12 storms common to the five trees (data of Watanabe and Ozeki 1964: 122, 128, 129). The greater interception efficiency and thus greater snow load per unit area of the 32-year-old tree are evident; otherwise there appears to be no relationship between snow load per m^2 and interception efficiency with age (Fig. 7.32). However, total snow load increased almost regularly with age. Two generalizations can be drawn. First, interception efficiency is not a simple function of age. Second, crown factors may influence interception per unit area and total snow load differently.

Of the crown attributes illustrated only HPA increased monotonically with age; the 32- and 57-year-old trees had equal VPA, both greater than the 43-year-old (Fig. 7.32). Because crown length was constant (4.8 m) and HPA increased monotonically with age so did crown volume and outside surface. The crown height to base ratio necessarily decreased monotonically with age (from 2.54 to 1.96). All but the 57-year-old tree were above the ratio suggested for rapid snow shedding on a flat plane in calm weather (Fig. 7.29), but recall that snow density declined linearly over slope angles of 20 to 60° (Eq. 7.23). The increase of total snow load with age could thus be any combination of increases in HPA, crown volume, and crown surface or the decrease in height to base

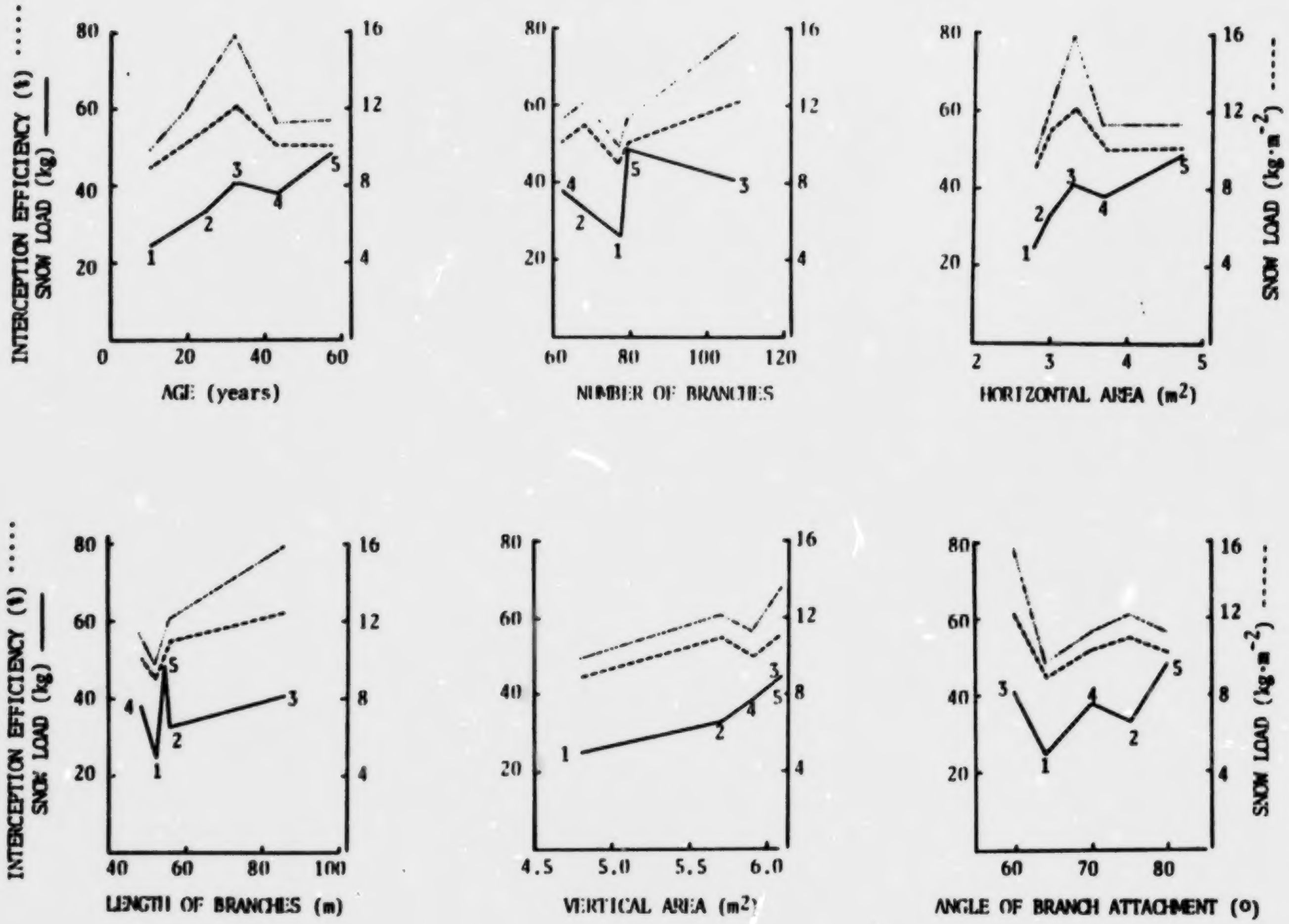


Figure 7.32 Interception efficiency, snow load, and snow load per unit area as a function of age, number of branches, horizontal area, length of branches, vertical area, and angle of branch attachment in five Cryptomeria japonica Kumasugi trees. Numbers refer to ages of trees: 1 = 10 years, 2 = 24 years, 3 = 32 years, 4 = 43 years, 5 = 57 years (data from Watanabe and Ozeki 1964: 122, 128, 129).

ratio.

The much higher interception efficiency of the 32-year-old tree could be associated with either its more numerous branches and their combined length, or the lower angle of branch attachment. None of these variables by themselves show a consistent trend with measurements of interception efficiency (Fig. 7.32). However, the combination of total branch length with crown volume to produce the crown density index indicated that the 32-year-old tree had by far the highest crown density (39% above the next highest).

The mean angle of primary branch attachment and subsequent branch flexibility also influence interception efficiency (Ch. 7.6.2 and 7.6.3). The data illustrated in Figure 7.33 are for the five experimental trees of Figure 7.32. Note that the 32-year-old tree (No. 3) exhibited the least depression of its branches with increasing snow load (Fig. 7.33a). As a result its vertical projected area was reduced only marginally (Fig. 7.33b) and its horizontal projected area was increased substantially under most snow loads experienced (Fig. 7.33c). The increase in horizontal area with increased snow load is inversely related to the initial branch angle. Only the youngest tree is out of order. It had the second lowest branch angle (Fig. 7.32), but its more flexible branches were depressed most by a given snow load (Fig. 7.33a) and its horizontal area was least increased by increasing snow load (Fig. 7.33c). These observations not only reveal a plausible explanation for the pattern of interception observed, but they

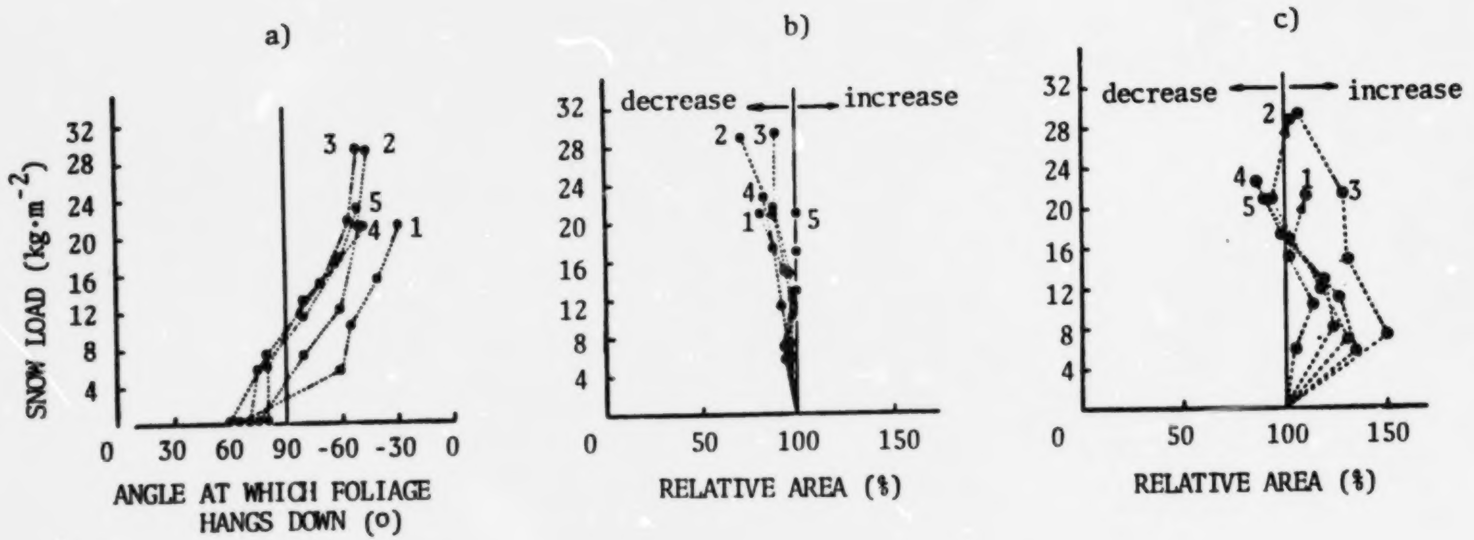


Figure 7.33 Change in the angle at which foliage hangs down and the crown area of five Kumasugi trees of different ages as a function of snow load (data of Watanabe and Ozeki 1964: 128,129). Numbers refer to ages of trees: 1 = 10 years, 2 = 24 years, 3 = 32 years, 4 = 43 years, 5 = 57 years.

a) Change in the angle of the foliage as a function of snow load.
b) Change in the vertical projected area of the crown.
c) Change in the horizontal projected area of the crown.

also reveal dangers inherent in examining only one crown attribute. For example, the youngest tree had the second highest index of crown density, but that was apparently countered by increased branch flexibility. The analyses further indicate the potential danger in utilizing crown measurements gathered in the absence of snow to predict interception. Flexible branches allow the crown to change shape with increasing snow load.

Data provided by Watanabe and Ozeki (1964) reveal a related reason why the 32-year-old tree was a more efficient interceptor. Not only did it contain many more branches (Fig. 7.32), but its branches were distributed unlike the other four trees, with a greater proportion of their total length concentrated in the lower half of the crown (Fig. 7.34). That vertical distribution of branches, together with their low angle of attachment and less flexible nature, would increase interception efficiency. Watanabe and Ozeki (1964) suggested that the fact that this tree was growing rapidly produced the crown characteristics so effective for interception. As both environment and age influence growth rate, the utility of age as a predictor of interception is reduced. It does seem likely, however, that when tree crowns contain a great density of branches they will be effective interceptors. The observation agrees with earlier work on model trees by JGFES (Fig. 7.5).

Only two of the crown attributes discussed could be estimated without great difficulty in the field: projected

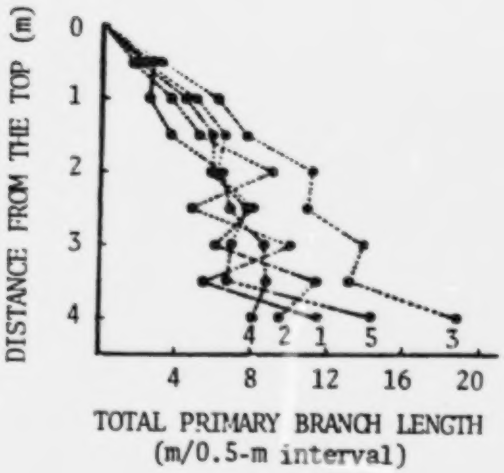


Figure 7.34 Vertical profile of branch density in five Kumasugi trees of different ages (from Watanabe and Ozeki 1964: 130). Numbers refer to ages of trees: 1 = 10 years, 2 = 24 years, 3 = 32 years, 4 = 43 years, 5 = 57 years.

horizontal and vertical area. The relationships between these and total snow load (SL, kg) were:

$$SL = 1.66 + 10.16\ HPA \qquad (7.27)$$
$$(r^2 = 0.80,\ SE = 4.38,\ P < 0.04)$$

$$SL = -44.55 + 14.3\ VPA \qquad (7.28)$$
$$(r^2 = 0.82,\ SE = 4.7,\ P \le 0.035)$$

If the effect of branch angle θ° and its implicit covariate, flexibility, are added to the relationship with vertical area, the relationship is much improved:

$$SL = 53.61 + 13.22\ VPA + 0.21\ \theta \qquad (7.29)$$
$$(r^2 = 0.95)$$

The second set of experiments involved modifying the crown of Kumasugi trees and observing the response to accumulated snow load. Modifications included chopping off the lower 1/3 or 1/2 of the branches (number, not length) and thinning the branches along the entire bole by 1/3 or 1/2 of their original number (Fig. 7.35). All trees were 30 years of age, near the apparent age of most efficient interception (Fig. 7.32). The data analyzed derive from 11 storms in January through March, 1962 (Watanabe and Ozeki 1964: 132-133). The storms were more intensive than those involved in the evaluation of age effects, ranging from 12.4 to 113.9 mm of snowfall (a mean of

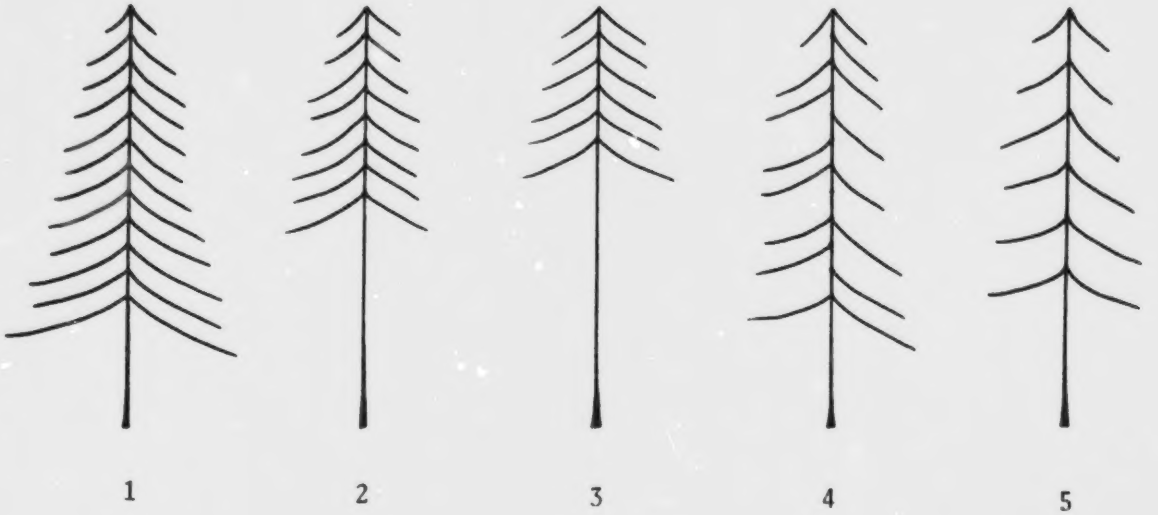


Figure 7.35 Schematic representation of five treatments applied to *Cryptomeria* crowns. Treatment 1 = control, 2 = 1/3 of lower branches removed, 3 = 1/2 of lower branches removed, 4 = 1/3 of branched thinned, 5 = 1/2 of branches thinned.

43.7 compared to 18.5 mm). The relative rankings of interception efficiency were consistent between treatments over the range of storm sizes: 1/3 branches thinned > 1/2 branches thinned > lower 1/3 branches removed > lower 1/2 branches removed (Fig. 7.36). The pattern of total snow load and load per unit area again differed and there was proportionately less difference in snow load ($kg \cdot m^{-2}$) than in interception efficiency (cf Figs. 7.36 and 7.37). Total snow loads on the control tree were as much as 40% higher than those on the treated trees, whereas snow load per unit area was seldom more than 20% higher (Fig. 7.38). The relative proportions of snow load attained by treated trees as a function of total load in the untreated tree were evaluated by regression analysis (Table 7.3).

Although treatments showed differences in the magnitude of snow intercepted, there were no apparent differences in the form of the relationships between snow load and snowfall between treatments (Figs. 7.36-7.38). We thus used the mean values for all 12 storms to evaluate the potential effects of crown attributes on interception. Total snow load responded far more to changes in crown attributes than did snow load per unit area (Fig. 7.39). Reductions in indices of crown area, such as HPA, VPA, and the number and length of branches, all reduced total snow load relative to the control tree. Because the two thinning treatments left the total length of the crown intact, three trees had crown lengths of 4.8 m (Fig. 7.39). These three trees, as well as the five of Figure 7.32, showed

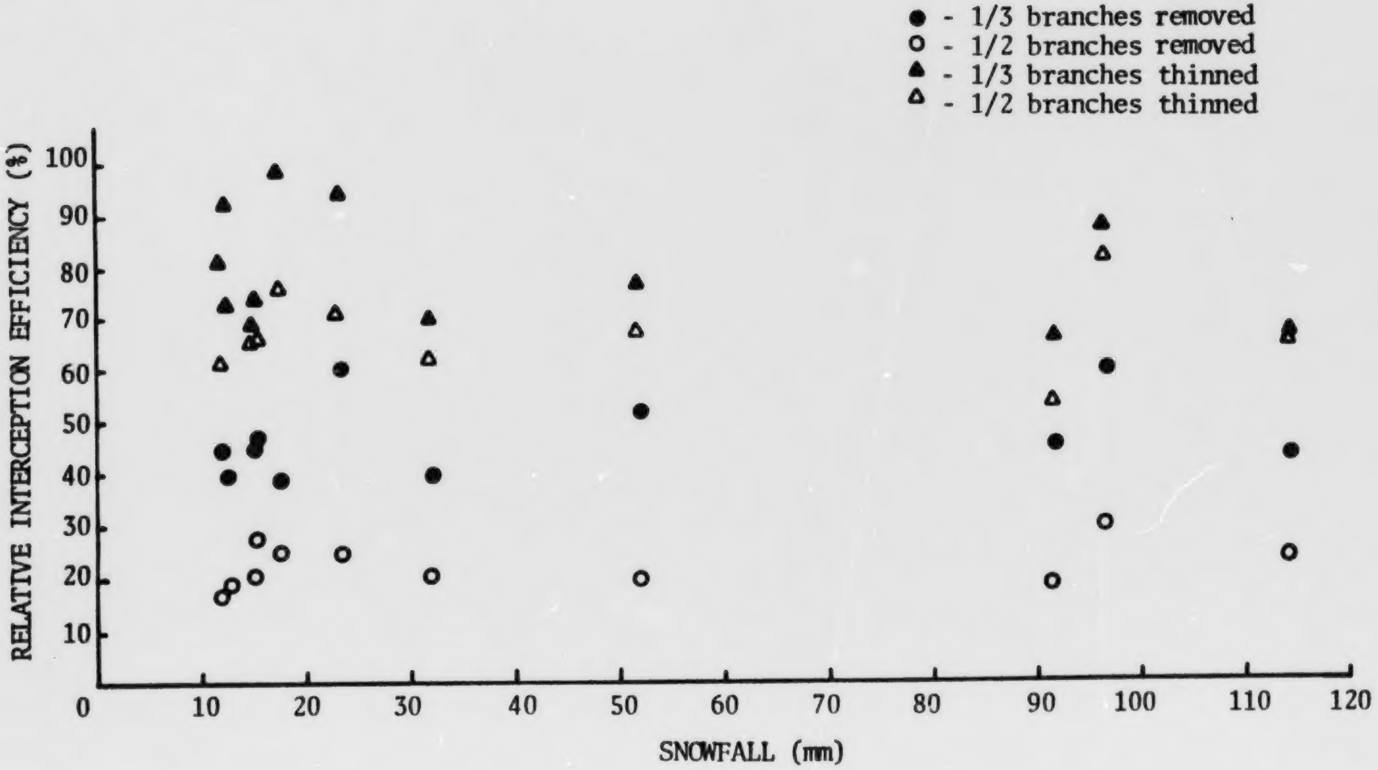


Figure 7.36 Relative interception efficiency (relative to untreated crown, %) of four treated Kumasugi crowns as a function of snowfall (from data of Watanabe and Ozeki 1964: 132, 133).

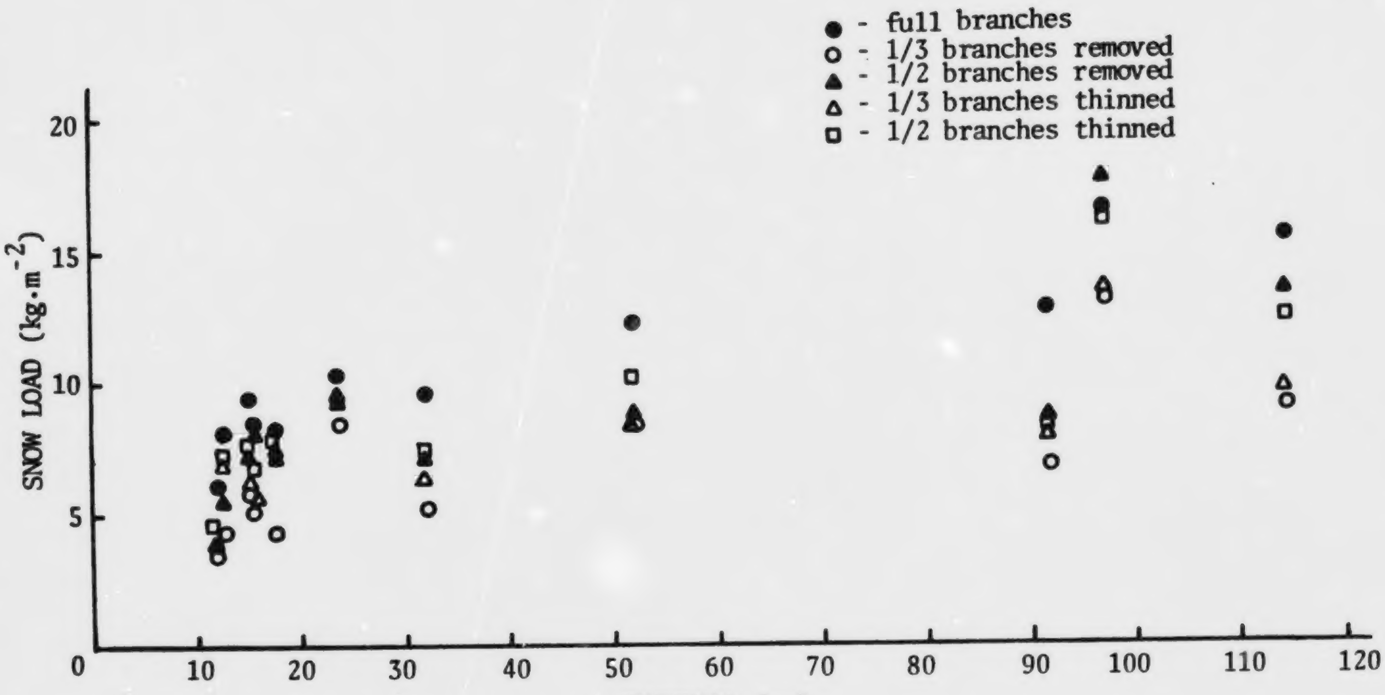


Figure 7.37 Crown snow load (kg·m^2) in four treated Kumasugi crowns as a function of total snowfall (from data of Watanabe and Ozeki 1965: 132, 133). The untreated crown is included for comparison.

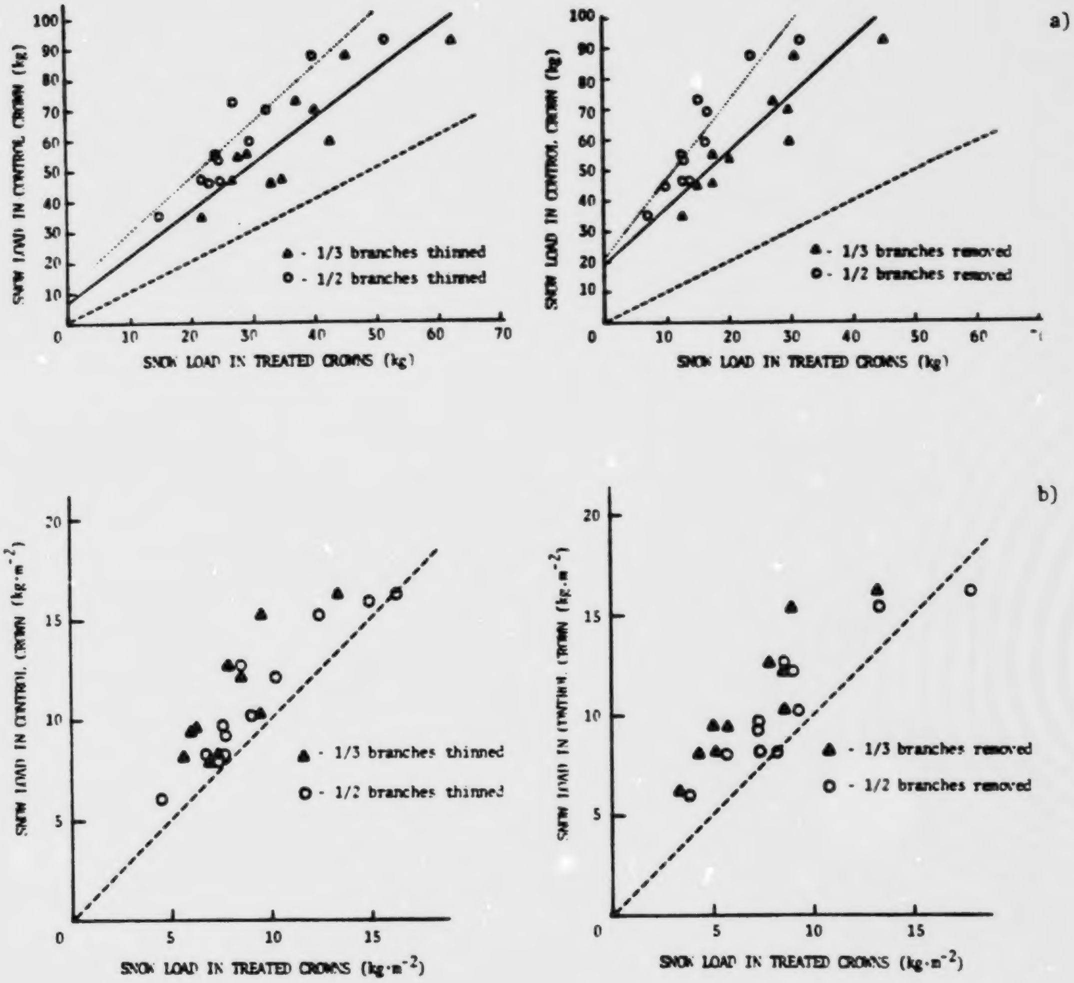


Figure 7.38 Crown snow load in the untreated crown relative to snow loads in treated crowns (data from Watanabe and Ozeki 1964: 132, 133).

a) Total snow load (kg). Dashed line represents a 1:1 ratio, other lines are regressions.

b) Snow load per unit area ($kg \cdot m^{-2}$). Dashed line represents a 1:1 ratio.

Table 7.3 Regression equations relating snow load on treated trees to snow load on the control tree (data of Watanabe and Ozeki 1964: 132-133). In regression equation y = control tree, x = treated tree.

Variable	Treatment	Equation	r^2	Sy.x	P(slope = 0)
Total snow load (kg)	1/3 branches thinned	$y = 8.2 + 1.4x$	0.77	9.28	0.0004
	1/2 branches thinned	$y = 11.99 + 1.7x$	0.86	7.19	0.0000
	1/3 branches removed	$y = 19.6 + 1.7x$	0.85	7.43	0.0000
	1/2 branches removed	$y = 21.5 + 2.5x$	0.84	7.80	0.0001
Snow load ($kg \cdot m^{-2}$)	1/3 branches thinned	$y = 1.5 + 1.2x$	0.76	1.64	0.0004
	1/2 branches thinned	$y = 2.1 + 0.96$	0.86	1.26	0.0000
	1/3 branches removed	$y = 3.5 + 1.04x$	0.85	1.31	0.0001
	1/2 branches removed	$y = 3.8 + 0.77x$	0.84	1.37	0.0001

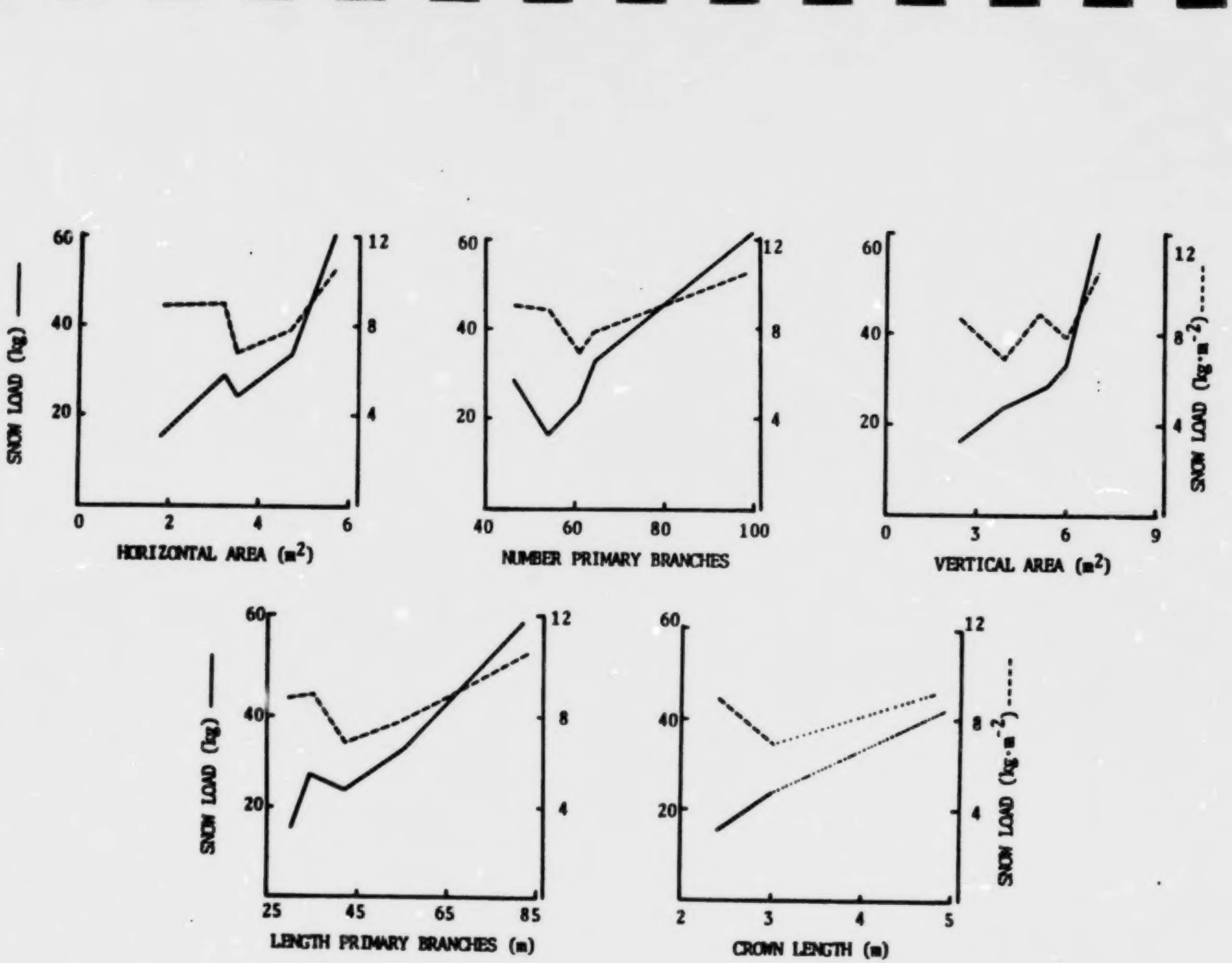


Figure 7.39 Total snow load and snow load per unit area as a function of several whole crown attributes horizontal area, vertical area, number of primary branches, length of primary branches, and crown length (data plotted are the means of 12 snowstorms from Watanabe and Ozeki 1964: 131, 132, 133). Sample size for crown length was only 3, see text.

marked differences in interception. Apparently crown length by itself is not a good predictor of interception.

Horizontal area could be accurately predicted by total branch length within the five trees in the experiment examining modified crowns ($HPA = 0.617 + 0.064\ BL$; $r^2 = 0.875$, $P < 0.02$), but vertical projected area could not ($r^2 = 0.685$). Among the five trees of different ages, HPA was unrelated to total branch length ($r^2 = 0.02$) yet remained an effective predictor of interception. We conclude that the apparent relationship between total branch length and total snow load (Fig. 7.39) is a result of branch length's fortuitous correlation with HPA.

We can evaluate the potential effect of the crown attributes considered by comparing the relative reduction in snow load associated with the relative reduction in each attribute (relative to control). Reductions in VPA and HPA both consistently reduced total snow load in broadly equivalent fashion (Fig. 7.40); reductions in the number and total length of primary branches had a more erratic effect (Fig. 7.39). We suggest that reductions in branch number and length modified total snow load only to the degree that they reduced the horizontal or vertical projected area.

Contrary to total snow load, relative snow load per unit area was increased by relative reductions in VPA or HPA exceeding 20%. Because older, lower branches were removed, relative reductions in branch lengths would change dramatically, branch numbers noticeably, but HPA and VPA would

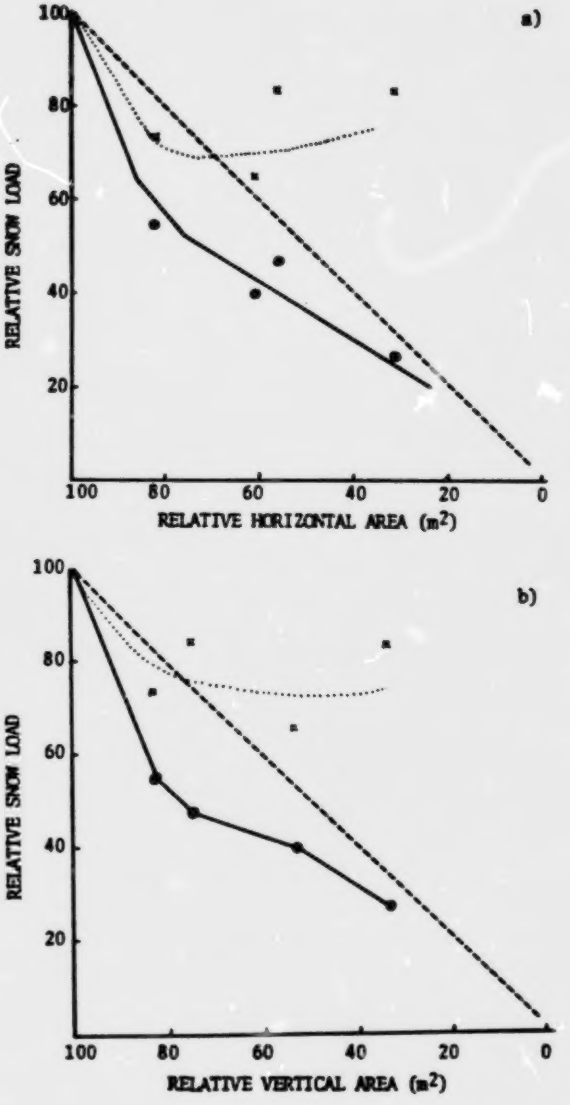


Figure 7.40 Change in snow loads of treated crowns as a function of the relative reduction in crown area (relative to control). Total snow loads (kg, —●—) and snow loads per unit area ($kg.m^{-2}$, ...x...) are means of 12 storms calculated from data of Watanabe and Ozeki 1964: 131, 132, 133.
a) Relative snow load as a function of relative horizontal area.
b) Relative snow load as a function of relative vertical area.

be reduced only to the extent that these lower branches extended beyond the 'edges' defined by upper branches. The upper, central and denser portion of the crown would provide more efficient interception than the 'edges'. Thus, the relative snow load per unit area would increase as the less efficient portions were removed. 'Thinning' would have had a different effect. Thinning probably increased the mean inter-whorl distance (Fig. 7.35). The work of JGFES 1952 on model crowns demonstrated that increased inter-whorl distance increased interception per unit area (Fig. 7.5). It appears that removal of the branches had a more direct effect on snow load per unit area than on total load. Because load per unit area increased with increasing branch removal (more efficient surfaces and greater inter-whorl distances remaining), the actual reduction in total load was decelerated.

Of the five potential crown attributes illustrated in Figure 7.39, two appear to have potential as predictor variables. These same variables were suggested in the analysis of potential age effects and have similar predictive capability for the data on treatment effects.

$$SL(kg) = -7.4 + 8.0\ VPA \qquad (7.30)$$

$$(r^2 = 0.80,\ SE = 8.8,\ P < 0.04)$$

$$SL(kg) = -6.7 + 10.3\ HPA \qquad (7.31)$$

$$(r^2 = 0.83,\ SE = 10.9,\ P \le 0.03)$$

With these particular data (modified crowns) the equations are more predictive in exponential form. Respective exponential counterparts to equations 7.30 and 7.31 are SL = $8.54\ e^{0\ 25}$ ($r^2 = 0.92$) and SL = $8.97\ e^{0\ 314}$ ($r^2 = 0.915$).

Summary evaluation.--The evaluation includes both the attributes measured directly by Watanabe and Ozeki (1964) and indices which we have derived from their and other workers' data. It is unclear whether the horizontal projected areas of Watanabe and Ozeki (1964) and Satterlund and Haupt (1967) are identical. We assume they are. Because total snow load increases with increasing snowfall (Figs. 7.12, 7.17, 7.31), and the snowfall varied during different experiments, we limited analyses to means of 16 storms in which snowfall was from 10 to 20 mm SWE. The range 10 to 20 mm encompassed the most storm events in each experiment. The two storms documented by Satterlund and Haupt (1967) were 9.4 mm SWE; they were included in the analyses unless noted.

For purposes of comparison the crown attributes are plotted versus common measurements of interception. This approach allows clearer illustration of how different crown attributes were related to the same amount of interception. The regression analyses treat interception as a function of crown attributes.

1. Age of tree.--Age probably has been used both because it can be assigned a value easily and because it might provide a surrogate for various aspects of crown form and flexibility.

The data reviewed indicate that: i) interception is better predicted by crown attributes that may be correlated with age than by age itself (Figs. 7.31 to 7.33); and ii) trees of the same age but different crown characteristics show differences in mean total snow load as great as 3.8 times (Fig. 7.40). We conclude that the value of age as a variable predicting interception is usually extremely limited. It may not be devoid of value, because it appears to correlate with branch flexibility (Fig. 7.33). Branch flexibility is important in interception, but difficult to measure.

2. Crown length.--Crown length by itself exhibited little correlation with interception efficiency, total snow load, or load per unit area (Fig. 7.39). Within the data reviewed eight trees had the same crown length (4.8 m), but intercepted snow very differently. These two variables (age and crown length) appear to be the only two that can be considered as strictly linear. Those variables which are meant to represent surface area proved to be better predictors of interception.

3. Horizontal and vertical projected area.--One would expect horizontal and vertical area to be closely related to total snow load, but not necessarily to load per unit area or interception efficiency. Indeed, that appears to be true (Fig. 7.41).

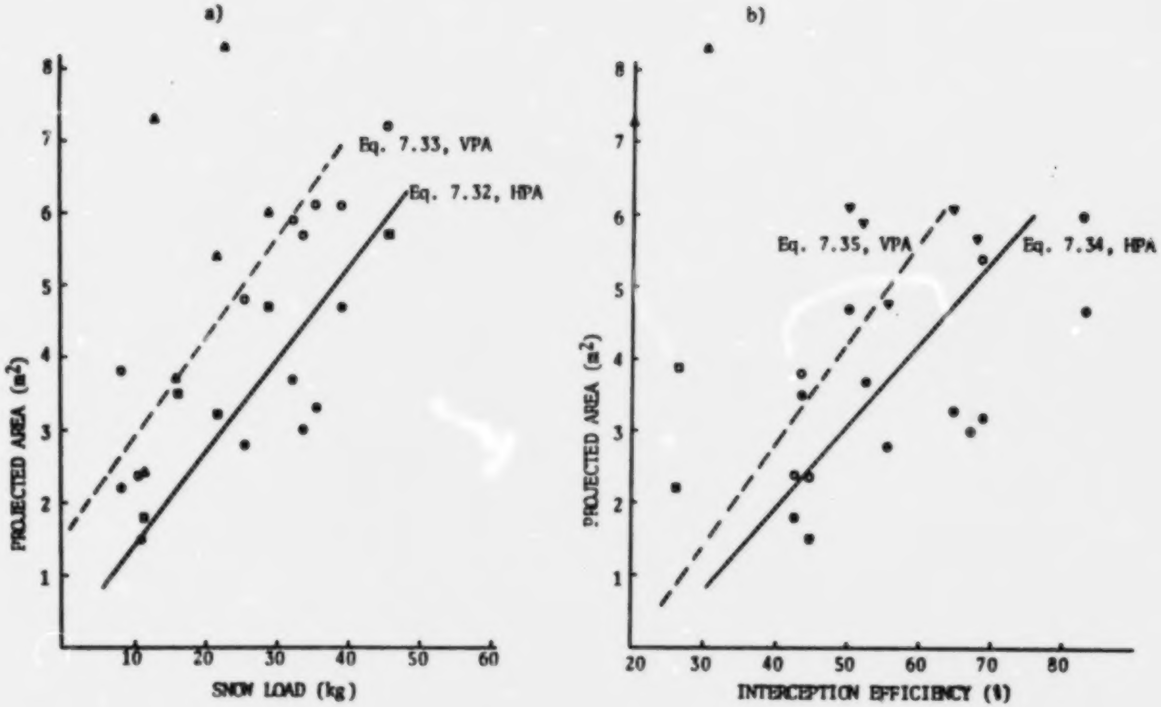


Figure 7.41 Relationships of horizontal and vertical projected area of the crown with total snow load and interception efficiency.
a) Total snow load (kg) and projected crown area. Horizontal area (m^2): ● = differently aged trees, Bokasugi and Kumasugi; ■ = modified crowns; ▲ = western white pine and Douglas fir. Vertical area (m^2): ○ = differently aged trees, Bokasugi and Kumasugi; △ = modified crowns.
b) Interception efficiency (%) and projected crown area. Horizontal area (m^2): ● = differently aged trees and modified crowns; ■ = Bokasugi and Kumasugi; ▲ = western white pine and Douglas fir. Vertical area (m^2): ○ = modified crowns, □ = Bokasugi and Kumasugi; ▽ = differently aged trees.

$$SL\ (kg) = -1.18 + 8.07\ HPA \tag{7.32}$$

$$(r^2 = 0.68,\ SE = 7.16,\ P < 0.001)$$

$$SL\ (kg) = -10.09 + 7.21\ VPA \tag{7.33}$$

$$(r^2 = 0.85,\ SE = 4.99,\ P < 0.0001)$$

$$IE\ (\%) = 27.9 + 8.45\ HPA \tag{7.34}$$

$$(r^2 = 0.31,\ SE = 13.7,\ P < 0.07)$$

$$IE\ (\%) = 20.26 + 7.15\ VPA \tag{7.35}$$

$$(r^2 = 0.43,\ SE = 12.5,\ P < 0.03)$$

Snow load per m^2 was unrelated to either HPA or VPA ($P > 0.3$).

The regression equations do not include the data of Satterlund and Haupt (1967). It is unclear whether the departure of their data from the pattern exhibited by data from Cryptomeria reflect differences between species or methodologies.

We expect the greater predictability of vertical area compared to horizontal area to be a function of the diagonal vector of snowfall. Unless snow is falling vertically, the vertical area plays a more important role in interception. Morey (1942) compared crown depth and closure, and concluded that a stand with more trees per acre and shallower canopy intercepts less snow than one with fewer trees and deep canopy.

It appears that the Japanese measurements of vertical projected area used a grid or other complex system, because their reported VPA cannot be calculated knowing only crown length and crown diameter (assuming HPA represents a circle). However, HPA accounted for 68% of the variation in total snowload and it is calculated from a simple, linear measurement - crown diameter. The apparent utility of crown diameter and the importance of vertical area, led us to examine a simple index: crown diameter x crown length (CI). The index predicted better than either HPA or VPA:

$$SL(kg) = -5.42 + 3.58\ CI \qquad (7.36)$$

$$(r^2 = 0.82,\ SE = 5.32,\ P < 0.0001)$$

$$IE(\%) = 23.15 + 3.74\ CI \qquad (7.37)$$

$$(r^2 = 0.48,\ SE = 11.9,\ P < 0.02)$$

The simplicity and the apparent predictive capability of the crown index suggest it may be useful for field use.

4. Crown surface.--Crown surface (CS, m^2) is also readily derived from only crown diameter and crown length (by assuming the crown is a cone). Like other explicit representations of surface it predicts total snow load better than interception efficiency (Fig. 7.42). As expected, there is no relationship with load per unit area:

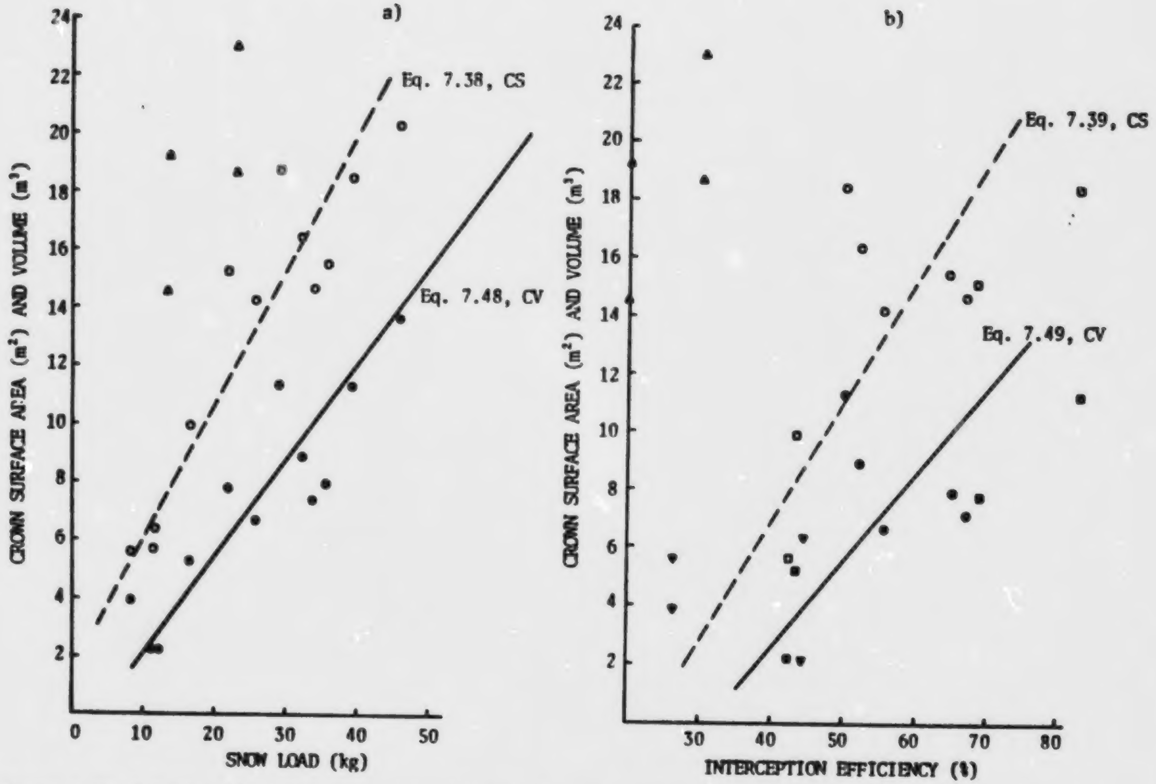


Figure 7.42 Relationships of surface area and volume of the crown with total snow load and interception efficiency.
a) Total snow load (kg) and surface area (m^2) or volume (m^3) of the crown assuming a conical shape. Crown volume: ● = differently aged trees, modified crowns, Bokasugi and Kumasugi; ▲ = western white pine and Douglas fir. Crown surface: ○ and □ = differently aged trees, modified crowns, Bokasugi and Kumasugi; △ = western white pine and Douglas fir.
b) Interception efficiency (kg) and surface area (m^2) or volume (m^3) of the crown assuming a conical shape. Crown volume: ● = differently aged trees; ■ = modified crowns, ▲ = western white pine and Douglas fir; ▼ = Bokasugi and Kumasugi. Crown surface: ○ = differently aged trees; □ = modified crowns; △ = western white pine and Douglas fir; ▽ = Bokasugi and Kumasugi.

$$SL(kg) = -3.10 + 2.15\ CS \tag{7.38}$$

$$(r^2 = 0.87,\ SE = 4.64,\ P < 0.00001)$$

$$IE(\%) = 23.6 + 2.4\ CS \tag{7.39}$$

$$(r^2 = 0.59,\ SE = 10.64,\ P < 0.006)$$

These regressions again exclude data of Satterlund and Haupt (1967).

Equations 7.38 and 7.39 are the best predictors of interception by individual trees we discovered (Fig. 7.42). By assuming a particular form to the crown (cone as compared to cylinder) they increase the proportion of variability statistically accounted for by 5% and 11% for total snow load and IE (%), respectively. The fact that these increments are small compared to the effectiveness of the simple index of diameter x length may simply be a product of the fact that both predictors are derived from length and diameter.

5. Height to base ratio.--The height to base ratio (H/B, crown length/crown diameter) should influence primarily the load per unit area. The index diameter x length (cylinder) is very nearly as predictive of total snow load as the index positing a conical form (Eq. 7.38 and 7.39), suggesting only a modest influence of form. Crown height to base ratio should not be an effective predictor of total snow load other than at extremes. However, load per unit area (kg/HPA) should increase with height:base ratio because of the diagonal vector

to snowfall. The greater the height per unit horizontal area, the more snow will be intercepted over that horizontal area. This reasoning is supported by the findings. Total snow load was unrelated to H/B, yet H/B was the best predictor found of snow load per unit area (Fig. 7.43):

$$SL(kg\cdot m^{-2}) = -0.37 + 3.88\ H/B \qquad (7.40)$$
$$(r^2 = 0.40,\ SE = 1.74,\ P < 0.028)$$

Possibly because this index is dimensionless, the data of Satterlund and Haupt (1967) fit the same pattern. When their data are included the relationship becomes:

$$SL(kg\cdot m^{-2}) = -3.54 + 5.30\ H/B \qquad (7.41)$$
$$(r^2 = 0.63,\ SE = 1.76,\ P < 0.0007)$$

Because load per unit projected horizontal area is a function of H/B, interception efficiency must also exhibit some relationship with H/B (although not necessarily as well expressed). Again using all data, the relationship is:

$$IE(\%) = -3.50 + 26.9\ H/B \qquad (7.42)$$
$$(r^2 = 0.39,\ SE = 14.9,\ P < 0.022)$$

Using only the Cryptomeria data this latter relationship was only poorly expressed ($r^2 = 0.11$).

It is noteworthy that the largest height to base ratios

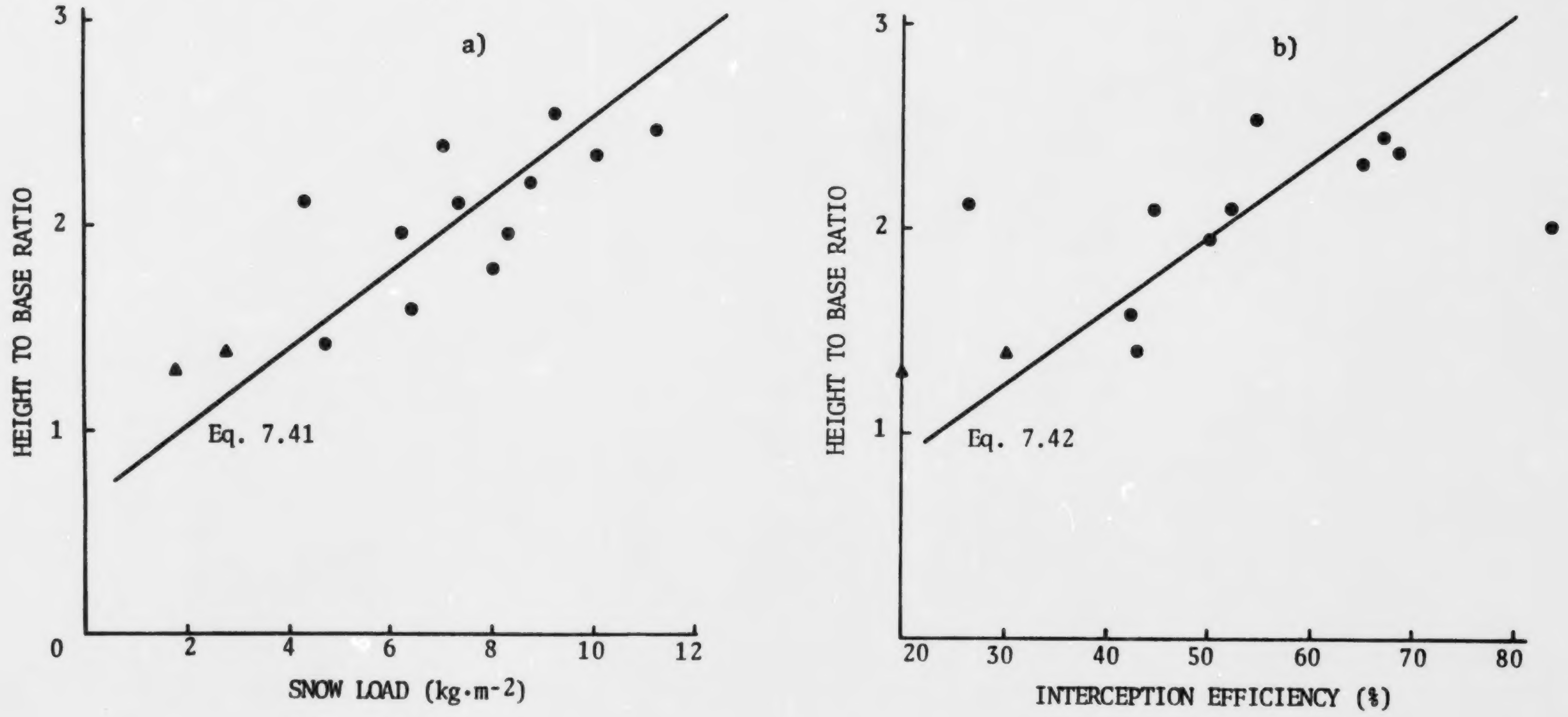


Figure 7.43 Relationships of height to base ratio with snow load per unit area and interception efficiency.
a) Snow load per unit area ($kg \cdot m^{-2}$): ● = differently aged trees, modified crowns, Bokasugi and Kumasugi; ▲ = western white pine and Douglas fir.
b) Interception efficiency (%): ● = differently aged trees, modified crowns, Bokasugi and Kumasugi; ▲ = western white pine and Douglas fir.

analysed exceeded only moderately the value of 2.15:1 which was suggested as being the point at which interception on sloping surfaces is markedly reduced (Fig. 7.29).

6. Branch number, length, and angle.--Number and total length of primary branches were highly correlated. We assume that length of the branches is more important to interception than their number. Greater total length of branches (BL,m) should indicate more surface area (greater total snow load) and possibly more surface area per unit HPA (greater load per unit area). Only the first trend was clearly evident (Fig. 7.44):

$$SL(kg) = -2.07 + 0.53\ BL \qquad (7.43)$$

$$(r^2 = 0.59,\ SE = 8.16,\ P < 0.0035)$$

$$SL(kg \cdot m^{-2}) = 4.31 + 0.06\ BL \qquad (7.44)$$

$$(r^2 = 0.27,\ SE = 1.9,\ P < 0.082)$$

Total branch length was a poorer predictor of total snow load than other indices of surface area. It was a better predictor of load per unit area, but the relationship was weakly expressed. Given that both total load and load per unit area were influenced, but differently, we expected no relationship between total branch length and interception efficiency. None was found (P > 0.25).

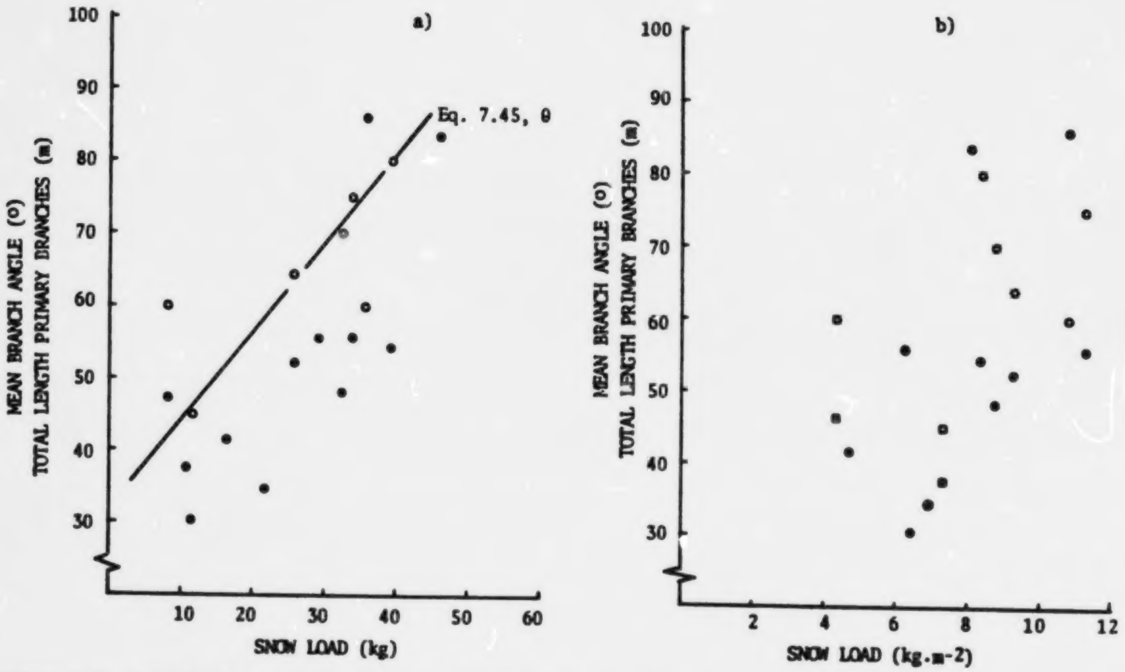


Figure 7.44 Relationships of mean angle and total length of primary branches with total snow load and snow load per unit area.
a) Total snow load with mean angle of primary branches from vertical in degrees (o) and total length of primary branches in meters (●).
b) Snow load per unit area (kg·m^{-2}). Total length of primary branches: ● = differently aged trees and modified crowns; ■ = Bokasugi and Kumasugi. Mean branch angle: o = differently aged trees; □ = Bokasugi and Kumasugi.

We anticipated that the effect of increasing branch angle (θ°) would be to increase all three measurements of interception. As branches approximated a more horizontal plane (90° displacement from trunk, see Fig.7.30), they should become more efficient interceptors. That in turn would increase both total snow load, and the load per projected horizontal area. Only the first effect was observed, with the same predictability as total branch length (Fig. 7.44):

$$SL(kg) = -26.07 + 0.81\,\theta \qquad (7.45)$$

$$(r^2 = 0.59,\ SE = 8.63,\ P < 0.043)$$

Neither load per m^2 nor interception efficiency were related to mean angle of the primary branches ($P > 0.4$).

We had data on branch angle for only seven trees. Within that sparse data base there are several potential reasons why mean angle was a poor predictor. They include: small sample size, variable flexibility of the branches under load, and sufficient surface available for these snowfalls (10-20 cm) that an additional increase in surface was superfluous. If the value for Kumasugi (which retained the least snow load per unit area) were eliminated, there would be an apparently steep relationship between branch angle and snow load per unit area (Fig. 7.44). Recall that Kumasugi is snow-adapted and has flexible branches under snow load (Fig. 7.30).

7. Indices of surface density.--Indices of crown surface were highly predictive of total snow load (Eqs. 7.33, 7.36, 7.38). We expected that an index of density would increase the predictablility by somehow approximating canopy completeness. We derived three indices by dividing the total length of primary branches by HPA, VPA, and crown surface (conical projection). Results were disappointing; values of r^2 never exceeded 0.11 for indices using HPA or VPA for any measure of interception. The index provided by dividing the total branch length by surface area of a cone (B/S) was better:

$$SL(kg) = 43.97 - 4.20\ B/S \qquad (7.46)$$

$$(r^2 = 0.32,\ SE = 10.50,\ P < 0.057)$$

$$IE(\%) = 78.48 - 5.54\ B/S \qquad (7.47)$$

$$(r^2 = 0.36,\ SE = 13.22,\ P < 0.051)$$

Both relationships are close to statistical significance ($P = 0.05$), but they function in a direction opposite to that which we predicted. We have only one plausible explanation. Five of the 12 study trees had their crown modified by removing branches. Relative reduction in branch length was greater than relative reduction in other crown attributes, leaving more efficient interceptor surfaces in a smaller area of crown (Fig. 7.40). Within these five crowns, component interceptor surfaces were less efficient the longer the total branch length; yet branch length and HPA were correlated. The net result is an apparent reduction in snow load using the B/S index.

Generally, the analyses of areal indices suggest that prediction of interception during moderate snowfalls is not enhanced by estimating only the length of all interceptor surfaces. They also suggest that if all branch diameters were measured, component surface areas would be predictive. When combined, the analyses suggest that branch density was seldom limiting in the crowns analyzed. Using current data simpler indices of area were more predictive. Because vertical projection is important, indices incorporating crown length were better predictors, although crown length alone was a poor predictor. Indices of total area were most predictive of total snow load, height:base ratios were most predictive of interception per unit area or interception efficiency. The indices of crown area also proved more predictive than indices of crown volume.

8. Crown volume.--If all interceptor surfaces were important, indices of volume encompassing all surfaces should be more predictive than indices of surface. The analyses involving branch lengths and angles suggest that under the snowfall conditions treated, consideration of all surfaces when they are measured only as length was unimportant. Whatever the condition, crown volume should predict total snow load better than snow load per unit area (Fig. 7.42). It

does:

$$SL(kg) = 3.63 + 3.0\ CV \qquad (7.48)$$

$$(r^2 = 0.81,\ SE = 5.54,\ P < 0.0001)$$

$$IE(\%) = 31.83 + 3.32\ CV \qquad (7.49)$$

$$(r^2 = 0.45,\ SE = 12.27,\ P < 0.024)$$

Crown volume in these estimates was computed as a cone. There was no relationship between crown volume and interception per unit area. Incorporating a third dimension to produce volume did not enhance the predictive ability beyond that of simple areal measurements. In fact, VPA and crown surface (conical) were marginally better predictors of total snow load. VPA was marginally less accurate in predicting interception efficiency.

9. Crown density.--Crown density was calculated by dividing total primary branch length by the volume of a cone. It was as ineffective at predicting total snow load (r^2 = 0.02) and interception efficiency ($r^2 = 0.31$, $P > 0.075$) as were other estimators incorporating branch density; probably for similar reasons. The major reason is probably that branch density was seldom limiting.

Summary.--The data available indicate that, during moderate snowfalls (10 to 20 cm), interception by single trees is best predicted by indices of area such as horizontal projected area or crown surface area. For total snow load on Cryptomeria trees, 82 to 87% of the variation could be accounted for by simple indices using crown diameter and length. Data for Douglas-fir and western white pine differed and may reflect inter-specific differences (e.g., branch and needle flexibility, angle of branch attachment, or branch density). The vertical component of crown area was important, and proved most effective when combined with crown diameter to estimate the surface of a cone ($r^2 = 0.87$), but even the simple index (crown diameter x crown length) accounted for 82% of the variation in total snow load.

Interception efficiency is largely determined by the less predictable load per unit area. It too was best predicted by the surface of a cone (Eq. 7.39; $r^2 = 0.59$). Snow load per unit area was most influenced by the relative importance of vertical and horizontal areas, as these were reflected by the height to base ratio. The relationship held for all species and accounted for 63% of the variation (Eq. 7.41).

It is unfortunate that no measurements of branch density or length were available for the North American species treated. It is possible that the dense branching pattern of the young Cryptomeria (similar to Thuja) generated the much greater rates of interception in Cryptomeria compared to North American species. The present analyses cannot eliminate the potential role of branch or surface density in the latter species.

7.6.5 Interspecific Differences.

The data reviewed above indicate that the species of tree can exercise an important influence on interception through its growth form, growth rate (inter-whorl distance), branch diameter and angle of attachment, branch and needle flexibility, and branch and needle roughness. Some of the differences in Figures 7.41 and 7.42 may be attributable to species differences. When growing in stands, the normal spacing of different species would further modify interception.

Many of the published observations on interspecific differences are anecdotal (Table 7.4). A somewhat more expansive treatment of most studies cited can be found in Shank and Bunnell (1982). The findings reviewed in Ch. 7.6.4 indicate that it should be possible, but tedious, to treat rigorously factors creating interspecific differences. The published statements we have summarized are largely qualitative. However, careful reading of Table 7.4 indicates that with one exception, observations under natural conditions confirm the experimental findings of JGFES (1952). The exception is the observation of Bühler (1886) that wide separation of whorls in Scots pine reduced snow storage, despite the long needles. That observation is contrary to findings of JGFES (1952) experiments on conical, stacked crowns (Fig. 7.5). Separation of boards in the JGFES experiments was not large, and the increased interception was

Table 7.4 Published observations on species-specific abilities to intercept snow.

Baldwin (1957)	Heavily loaded branches of fir, spruce, and hemlock droop and let snow slide off, while the stiffer branches of pine continue to hold it. In large storms, the flexibility of pine needles may be countered by the stiffness of pine branches.
Bennett (1959)	Hardwoods are more vulnerable to ice-loading than are conifers. Conifers have more resilient branches, smaller crowns, and a mechanically stronger branching habit. Even brittle-wooded trees survive heavy storms if they have coarse branching.
Bühler (1886) Germany	Breakage occurred when snow catch amounted to 26 mm in hardwoods and 46 mm in conifers. Hardwoods have more vulnerable branching habits. Wide separation of branch whorls reduces snow storage in Scots pines. Snow lies in the interior of hardwoods but on the upper whorls and ends of conifer branches.
Connaughton (1935) Idaho	A ponderosa pine reproduction plot averaged 5.4 per cent snow interception while a virgin ponderosa pine forest intercepted 24.5 per cent.
Delfs (1955 in Miller, 1964) Europe	Snow interception was least in the Kammfichte (comb-spruce) variety and most in the Burstenfichte (brush-spruce) variety that has side branches rising sharply upward.
Eastman (1978) Northern B.C.	Snow depths at open or deciduous sites, partially cutover sites, and conifers varied in the approximate ratio 100:75:50.
Fowler and Berndt (1971) Washington	Branches of alpine fir and lodgepole pine showed no differences in rime collection.
Gay (1958) Australia	Alpine-ash (Eucalyptus delegatensis)intercepts little snow because its branching habit is erect, its foliage is open, and the twigs are flexible. Snowgum eucalyptus (E. pauciflora) has heavy leaves, stiff twigs, and petioles which mat down as snow piles on them.
Kienholz (1940) Connecticut	Pine and hemlock stands intercept most light snowfalls but not heavy ones. They intercept more than hardwoods.
Kittredge (1953) California	Largest average interception was 27.5% occurring in an old growth sugar pine/ponderosa pine stand. Snow interception is further generalized by species as follows: immature red fir 20%, immature white fir 25%, pole size and mature ponderosa pine 15%, mature white fir 12%.
Klintsov (1958) USSR	The difference in snow load between pine and spruce (8 and 6 mm respectively) is considered to be due to the greater rigidity of pine branches.

Table 7.4 (continued)

Love (1955)	Insect-killed lodgepole pine and Englemann spruce that had lost needles and fine branches intercepted less snow than live trees, but remaining branches still intercepted some snowfall.
Lull and Rushmore (1961) New York	Time lapse photography showed white pine needles accumulated snow only at the fascicles until they were bent into a platform. Balsam fir was the best collector of snow because of stiff branches and persistent needles on the branches. Hemlock was the worst species studied because of feathery and flexible needles. Red spruce was intermediate in snow-holding capacity.
Maule (1934) Connecticut	White pine have longer and softer needles and more flexible branches than Norway spruce or red pine and allow more snow to reach the ground. Norway spruce, because of short branches and stiff needles, holds a great amount of snow. Hemlock have as little effect on retarding snow melt as have hardwood species (possibly due to uneven crown formation). Red pine was equal to white pine in interception ability. Ranking by accumulation: open>hardwood>hemlock>white pine>red pine>Norway spruce.
Pierce et al. (1958) Eastern U.S.	More snow accumulation in hardwood stands than in conifer stands.
Rosenfeld (1944) Germany	Ratio of branch wood to stem size is important. Spruce stands experiencing worst snow damage were in age classes in which dense stocking produced slender, high-crowned trees (large height to base ratio).
Satterlund and Haupt (1970) Northern Idaho	Interception by Douglas-fir and western white pine trees was 32.1% and 35.5% respectively. No difference in total snow catch or in the mode and timing of snow loss from crowns were evident by species.
Singer (1979) Montana	Snow depths in the open were 80 cm, in Douglas-fir stands 12 cm, and in spruce forest 40 cm.
Smith (1974)	Fir has flexible boughs causing them to droop with snow load and drip water into a narrow ring 3-6 feet from the stem. In contrast, lodgepole pine had stiff boughs and melt-water drops from point of storage. Therefore, SWE on the ground in fir stands is only 60-70% of the open whereas in lodgepole stands, it is 93-97% of the open.
Trimble (1956) Vermont	Snow interception is generalized by species as follows: northern hardwoods 10%, aspen-birch 7%, spruce or spruce-fir 35%, white pine 25%, hemlock 25%, red pine 30%.
Watanabe and Ozeki (1964) Japan	Snow loads per unit area of 2 *Cryptomeria* varieties differed by 1.8 times on average, a function of differences in angle of branch attachment and foliage flexibility. The more flexible crowns contained least snow.

Table 7.4 (continued)

Weitzman and Bay (1959) Minnesota	Snow accumulation was greatest in aspen followed in order by red pine thinned to 60, 100, and 140 ft^2 acre-1.
Wellington (1950)	Hardwoods intercept less snow than conifers because they vibrate more in the wind; not because of less surface area.
Zon (1927) USSR	Broadleaf forests (birch 35-70 years and oak 25-90 years) will contain about 41% more snow per unit area than will pure pine forest (20-90 years), 60% more than a pure spruce forest, and 50% more than coniferous forests in general.

assumed to reflect transport by wind into the interior. The optimum separation of whorls for interception likely varies with wind speed and angle of branch attachment. Bühler's observations are consistent with the findings of Watanabe and Ozeki (1964) if we assume their thinning experiments effectively increased mean inter-whorl distance (Figs. 7.35 and 7.38).

Other observations summarized (Table 7.4) are congruent with findings of the few quantitative studies treating tree morphology.

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8. FACTORS INFLUENCING INTERCEPTION IN STANDS

Chapter 7 documented that a large number of interacting factors influence interception by single trees. Notably temperature and wind speed interact with the size, shape, structure, and physical or mechanical characteristics of tree crowns. In stands the phenomena of interception and accumulated snow load are more complex because individual crowns are not identical and are not uniformly distributed. Kittredge (1953) found that the percentage interception beneath the crowns of individual trees was generally higher than the average for the entire stand (Ch. 8.1.1). Furthermore, if a forest stand is a mosaic of mixed species groups, then stand interception becomes the integral of species-specific abilities to intercept snow (Table 7.4). The stand itself influences abiotic variables, particularly air motion (Ch. 4). The individual crown attributes interact together and with abiotic factors in a complex manner.

A stand measurement necessarily takes longer to acquire than do those from individual trees. Reported measurements are integrated over a longer period of time and over a larger, invariably somewhat heterogeneous, area. As a result, the relative contributions of individual variables or processes cannot be separated clearly. Phenomena such as the transport of intercepted snow from trees during snow storms (e.g., Fig. 7.10) still occur but are integrated inseparably in the broad measurements acquired.

We have attempted to reduce the ambiguity by concentrating on measurements from individual storms. Nevertheless the variables evaluated remain aggregates or surrogates of the underlying processes. Two potentially overriding factors are evaluated first: storm size and elevation. The evaluation consists of comparing empirical observations with predictions derived from single tree measurements (Ch. 8.1). Examination of the influence of tree characteristics on interception efficiency is flawed because we have had to incorporate some measurements of snowpack beyond the duration of an individual storm, and could treat only one crown measurement adequately. The empirical summaries are informative but cannot separate individual processes (Ch. 8.2). Because stands are spatially heterogeneous we also examine the effects of forest openings on apparent interception. General principles of air motion in stands are discussed before observed patterns are examined (Ch. 8.3). The examination of observed patterns has had to rely largely on measurements of snowpack and individual processes could not be separated (Ch. 8.4).

8.1 Storm Size and Elevation

The data reviewed for single trees suggest a large number of factors influence efficiency of interception. For the reasons noted above, all of these cannot be examined independently and aggregate or surrogate variables must be employed. Two of these are storm size and elevation. Efficiency of interception and snow load of a single tree (Figs. 7.13-7.15) or even a single board (Figs. 7.20 and 7.21) changed with storm size in some sigmoidal pattern. We would expect that interception by stands should follow a similar pattern. Temperature also has strong effects on interception efficiency (Ch. 7.2) but temperatures are infrequently recorded during studies of interception. We have used elevation as a surrogate for temperature. On the basis of data from individual trees or single boards (Figs. 7.2 and 7.3) we would expect snow at lower elevations to be intercepted more efficiently because it is more likely to be warm and wet.

8.1.1 Storm Size

Table 8.1 summarizes available data on intercepted snow in stands during single snow storms. Perhaps the best data set is that of Fitzharris (1975) who compared SWE's in the open and within stands under individual trees at twelve elevations during 80 storms. Crown cover in sampled stands at each elevation was estimated by random sampling ($n = 25$) using ocular estimation as the crown closure index. Representative crown closures for data of Fitzharris are included in Table 8.1. Figures 8.1 to 8.6 present scatter plots of apparent snow load vs snowfall. As expected there is significantly more scatter than in similar plots made repeatedly for the

Table 8.1 Snow interception in stands during individual storms.

Source and Location	Date	Stand type	Canopy cover(%)	Snow water equivalent (mm)			Apparent Interception (%)	Remarks
				Load	Under canopy	Open		
Munns (1921) in U.S. Army (1956); California		Jack pine	80	0.03	0.00	0.03	100	Mean of several storms. Refers to mixed rain and snow converted from SWE (inches).
				0.04-0.09	0.01-0.04	0.05-0.13	73	
				0.11-0.24	0.04-0.06	0.15-0.25	76	
				0.11-0.30	0.17-0.46	0.28-0.76	39	
				0.33	0.46-0.74	0.79-1.27	42	
					0.96-1.86	1.30-2.51	26	
					2.00-3.96	2.57-5.08	22	
					3.88	5.11+	24	
Maule (1934) Connecticut	12.10	Hardwood; 3 age classes (1-20, 20-40, 40-60yrs); 6.1-19.8 m in height	100	0.51	2.54	3.05	17	Snow values extracted from his Figure 1. Snow from individual storms was measured. Snow measured in inches depth and transformed here on the basis of a density of 0.1 gm·cm^{-3}.
	12.13			0.25	3.81	4.06	7	
	12.17			0.00	13.21	13.21	0	
	01.29			0.00	2.03	2.03	0	
	02.04			2.54	3.05	5.59	45	
	02.11			0.00	17.27	17.78	3	
	12.10	Red pine; 7.3 m in height; 11-20 yrs	9-14	2.03	1.02	3.05	67	
	12.13			2.54	1.52	4.06	63	
	12.17			7.11	6.10	13.21	54	
	01.29			0.51	1.52	2.03	25	
	02.04			3.05	2.54	5.59	55	
	02.11			4.32	13.46	17.78	24	
	12.10	Norway spruce; 9.1 m in height; 11-20 yrs	6-7	2.29	0.76	3.05	75	
	12.13			2.79	1.27	4.06	69	
	12.17			8.64	4.57	13.21	65	
	01.29			1.01	1.02	2.03	50	
	02.04			4.32	1.27	5.59	77	
	02.11			8.13	9.65	17.78	46	
	12.10	White pine; 7.9 m in height; 11-20 yrs	5	1.53	1.52	3.05	50	
	12.13			3.04	1.02	4.06	75	
	12.17			6.86	6.35	13.21	52	
	01.21			0.76	1.27	2.03	38	
	02.04			3.05	2.54	5.59	55	
	02.11			2.54	15.24	17.78	14	

Table 8.1 (continued)

	12.10	Hemlock; 14.6-21.3 m in height; uneven age	12-13	1.14	1.91	3.05	38	
	12.13			2.15	1.91	4.06	53	
	12.17			6.86	6.35	13.21	52	
	01.29			0.76	1.27	2.03	38	
	02.04			3.05	2.54	5.59	55	
	02.11			4.32	13.46	17.78	24	
Johnson (1942) Colorado		Ponderosa pine		0.76-1.27				13 rainstorms analyzed. It is suggested on no evidence that maximal rain load is equal to maximal snow load.
Morey (1942) Vermont	04.11	Hardwood; fully stocked 60-yr-old		2.03		20.83	10	Measured after snow had blown off.
	04.11	Spruce; 30-yr-old		9.14		20.83	44	
Kittredge (1953) California	Winters of 1934-38 and 1940-41	White fir; mature 140 yrs	51	0.43	0.57	1.00	43	110 storms measured; no upper limit to interception although some cryptic comments about y-intercept being "snow storage". Data are computed from his regression equations p.9. Canopy cover is average within 6.1 meters of station.
				0.48	1.52	2.00	24	
				0.63	4.37	5.00	13	
				1.31	13.87	15.00	7	
		Ponderosa pine; mature	35	0.33	0.67	1.00	33	
				0.43	1.57	2.00	21	
				0.73	4.27	5.00	15	
				1.73	13.27	15.00	12	
		Ponderosa pine; 4.27 m	40	0.14	0.86	1.00	14	
				0.25	1.76	2.00	12	
				0.58	4.42	5.00	12	
				1.68	13.32	15.00	11	
		Red fir	75	0.89	0.11	1.00	89	
				1.02	0.98	2.00	51	
				1.41	3.59	5.00	28	
				2.71	12.29	15.00	18	
		White fir; pole size	70	0.83	0.17	1.00	83	
				1.00	1.00	2.00	50	
				1.51	3.99	5.00	30	
				3.21	11.79	15.00	21	

Table 8.1 (continued)

		Mixed conifer; cutover	55	0.88	0.13	1.00	87	
				1.12	0.88	2.00	56	
				1.84	3.16	5.00	37	
				4.24	10.76	15.00	28	
		Sugar/ ponderosa pine	62	0.53	0.47	1.00	53	
				0.81	1.19	2.00	41	
				1.65	3.35	5.00	33	
				4.45	10.54	15.00	30	
Strobel (1978) Alp mountains	01.06	uneven-aged coniferous; 29.3 m^2/ha	61	1.15	1.29	2.44	47	Data are for individual storms.
	01.14			0.45	0.86	1.31	34	
	01.16			0.88	0.53	1.41	62	
	01.18			1.77	2.20	3.97	45	
	01.22			1.39	4.14	5.53	25	
	01.28			2.05	8.02	10.07	20	
	01.06	uneven-aged coniferous; 75.1 m^2/ha	86	1.25	1.16	2.41	52	
	01.14			0.90	0.67	1.57	57	
	01.16			0.85	0.48	1.53	69	
	01.18			1.76	2.27	4.03	44	
	01.22			2.45	3.70	6.15	40	
	01.28			3.33	7.47	10.80	31	
Rowe and Hendrix (1951) California	1940-1946	Ponderosa pine; 65-70 yr-old; 1450 trees/ha; 6.7-33.8 m in height; elevation 1005 m	40	0.13	1.27	1.40	9	Data are for storms (> 1.0 cm SWE) in which ≥ 50% of precipitation fell as snow. No evidence of upper limit to interception.
				0.25	1.40	1.65	15	
				0.38	1.65	2.03	19	
				0.39	1.90	2.29	17	
				0.50	2.29	2.79	18	
				0.76	2.16	2.92	26	
				0.25	2.67	2.92	9	
				0.89	2.29	3.18	28	
				0.38	2.92	3.30	31	
				0.51	3.05	3.56	14	
				0.38	3.30	3.68	10	
				0.26	3.68	3.94	7	
				0.51	3.81	4.32	12	
				0.38	4.06	4.44	9	
				0.76	3.81	4.57	17	
				0.77	4.06	4.83	16	
				0.38	4.57	4.95	8	
				1.02	4.57	5.59	18	

Table 8.1 (concluded)

				0.64	5.08	5.72	11	
				0.76	6.22	6.98	11	
				0.77	6.98	7.75	10	
				1.01	7.37	8.38	12	
				1.26	9.14	10.40	12	
				1.02	10.41	11.43	9	
				2.16	13.46	15.62	14	
				1.01	15.88	16.89	6	
				2.67	20.70	23.37	11	
				3.68	26.42	30.10	12	
Fitzharris (1975) Coastal B.C.	1969-1971	Mixed conifer; mean height 13 m; elevation 590 m	51	0.10	0.10	0.30	67	82 individual storms were measured. Data here represent a subset of his data chosen for a range of canopy closures and snow storm sizes.
				1.90	1.40	3.30	18	
				3.50	3.90	7.40	7	
				0.10	0.00	0.10	100	
				0.20	0.00	0.20	100	
				0.30	0.00	0.30	100	
		Mixed conifer; mean height 12 m; elevation 710 m	91	0.03	0.00	0.30	100	
				3.50	1.00	4.50	98	
				1.50	7.80	9.30	16	
				0.20	0.00	0.20	100	
				0.80	0.20	1.00	80	
				2.00	0.60	2.60	79	
		Mixed conifer; elevation 790 m	71	0.30	0.20	0.50	60	
				2.90	7.70	10.60	27	
				2.55	2.35	4.90	52	
				0.20	0.00	0.20	100	
				1.10	0.60	1.70	65	
				2.9	0.70	3.60	80	
		Mixed conifer; mean height 10 m; elevation 1060 m	29	0.30	0.00	0.30	100	
				0.80	6.60	7.40	11	
				5.00	9.90	14.90	34	
				0.30	0.00	0.30	100	
				3.00	1.00	4.00	75	
				6.40	1.70	8.10	79	

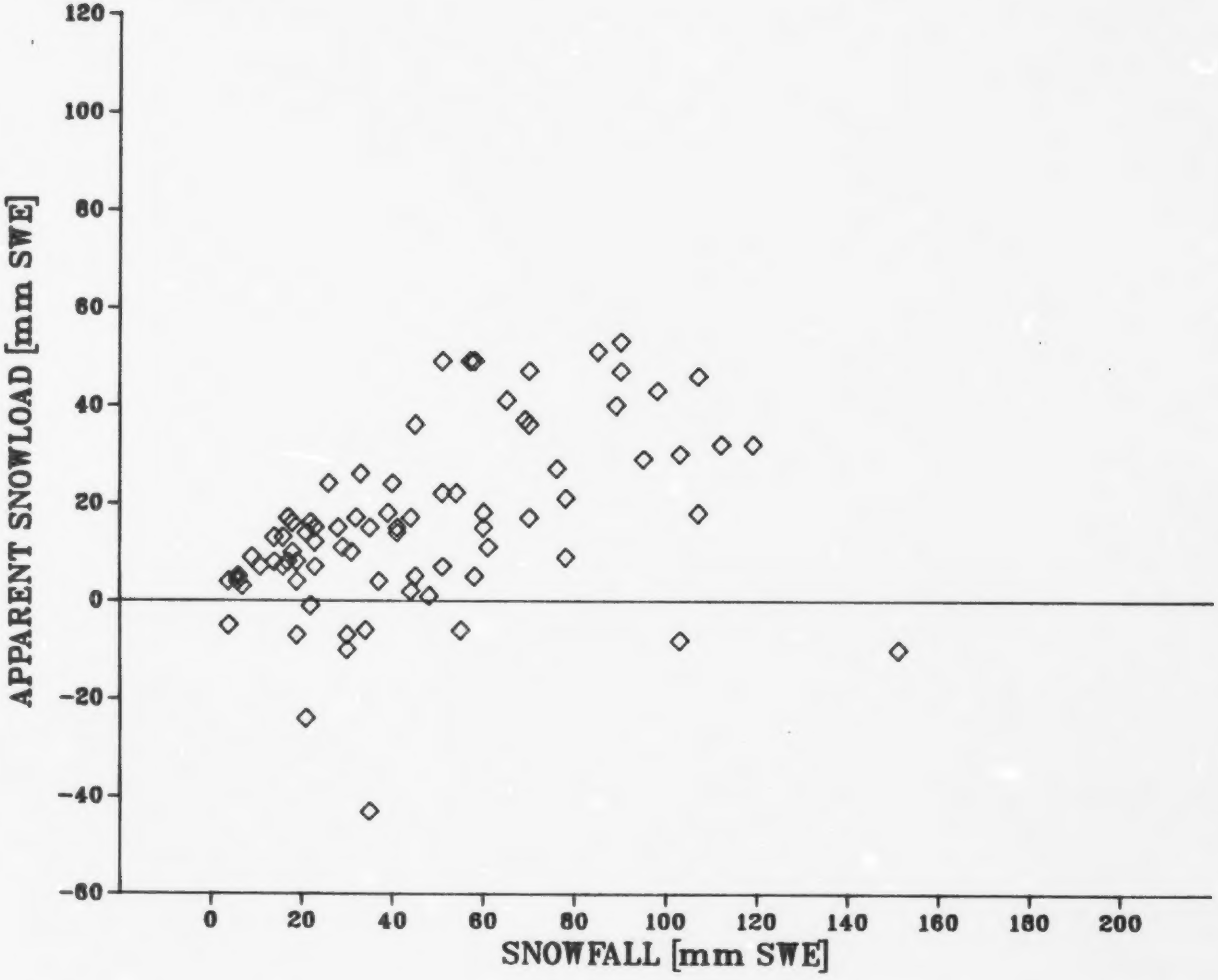

Figure 8.1 Apparent snow load in trees during individual storms at 1260 m on Mt. Seymour (data of Fitzharris 1975: Appendix G).

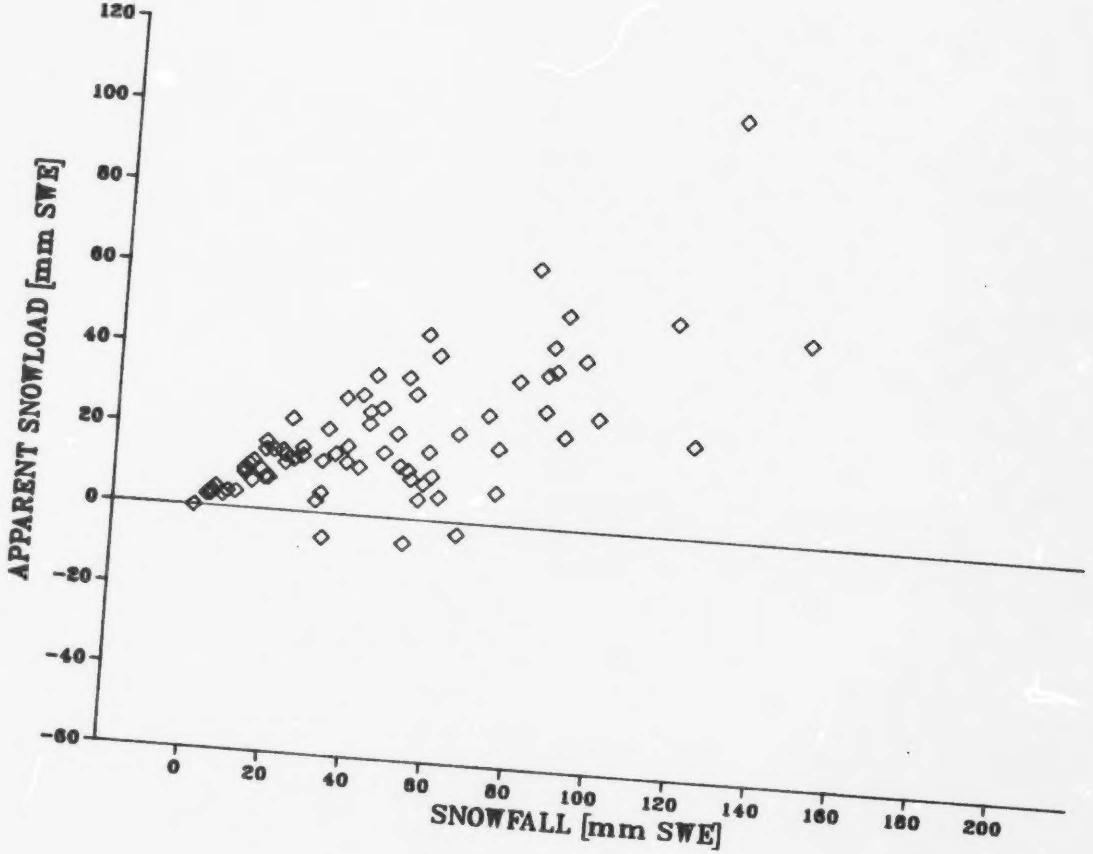

Figure 8.2 Apparent snow load in trees during individual storms at 1060 m on Mt. Seymour (data of Fitzharris 1975: Appendix G).

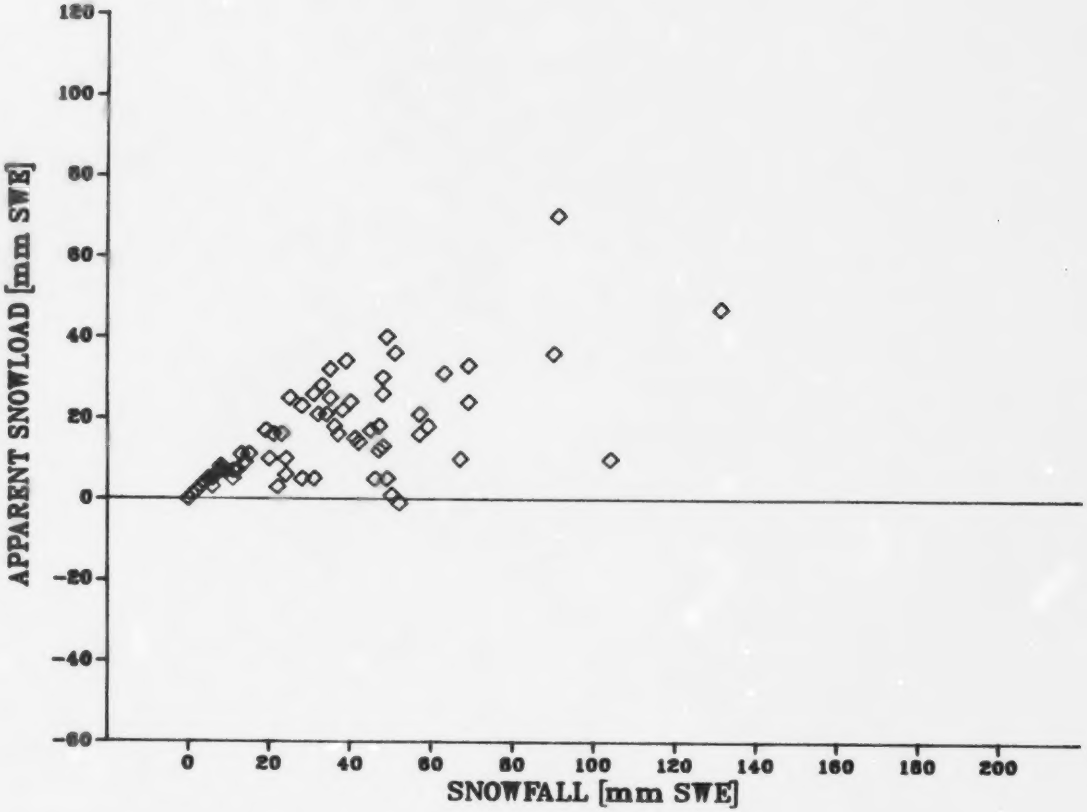

Figure 8.3 Apparent snow load in trees during individual storms at 970 m on Mt. Seymour (data of Fitzharris 1975: Appendix G).

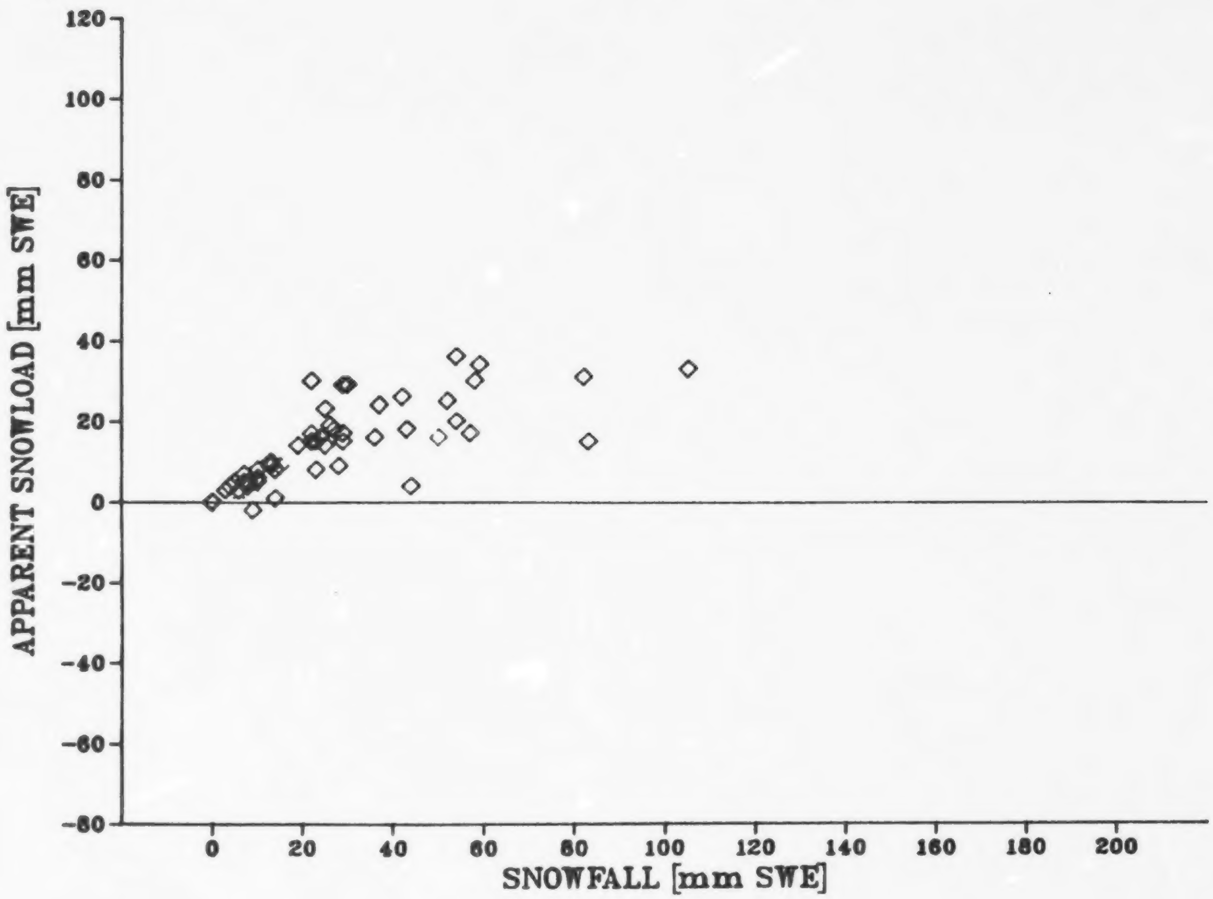

Figure 8.4 Apparent snow load in trees during individual storms at 870 m on Mt. Seymour (data of Fitzharris 1975: Appendix G).

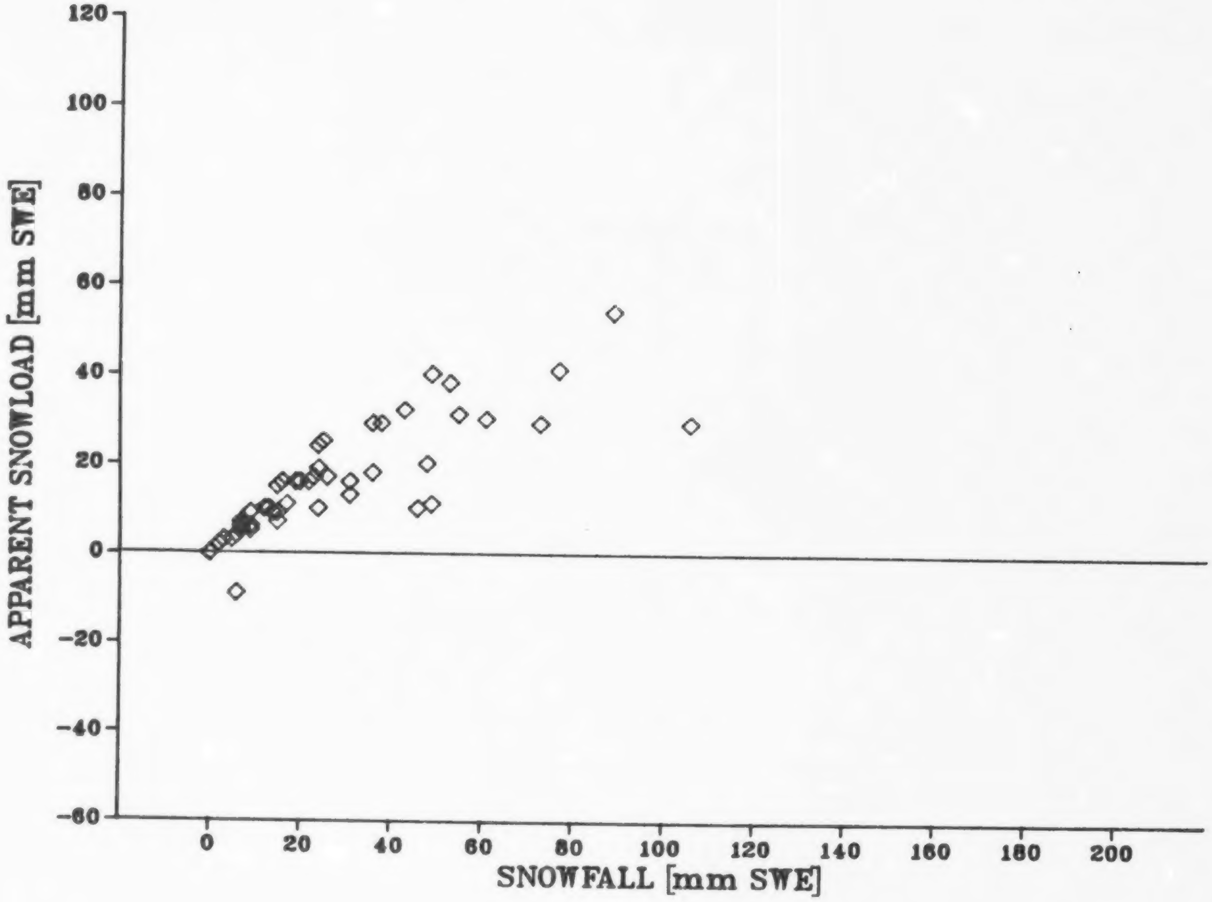

Figure 8.5 Apparent snow load in trees during individual storms at 790 m on Mt. Seymour (data of Fitzharris 1975: Appendix G).

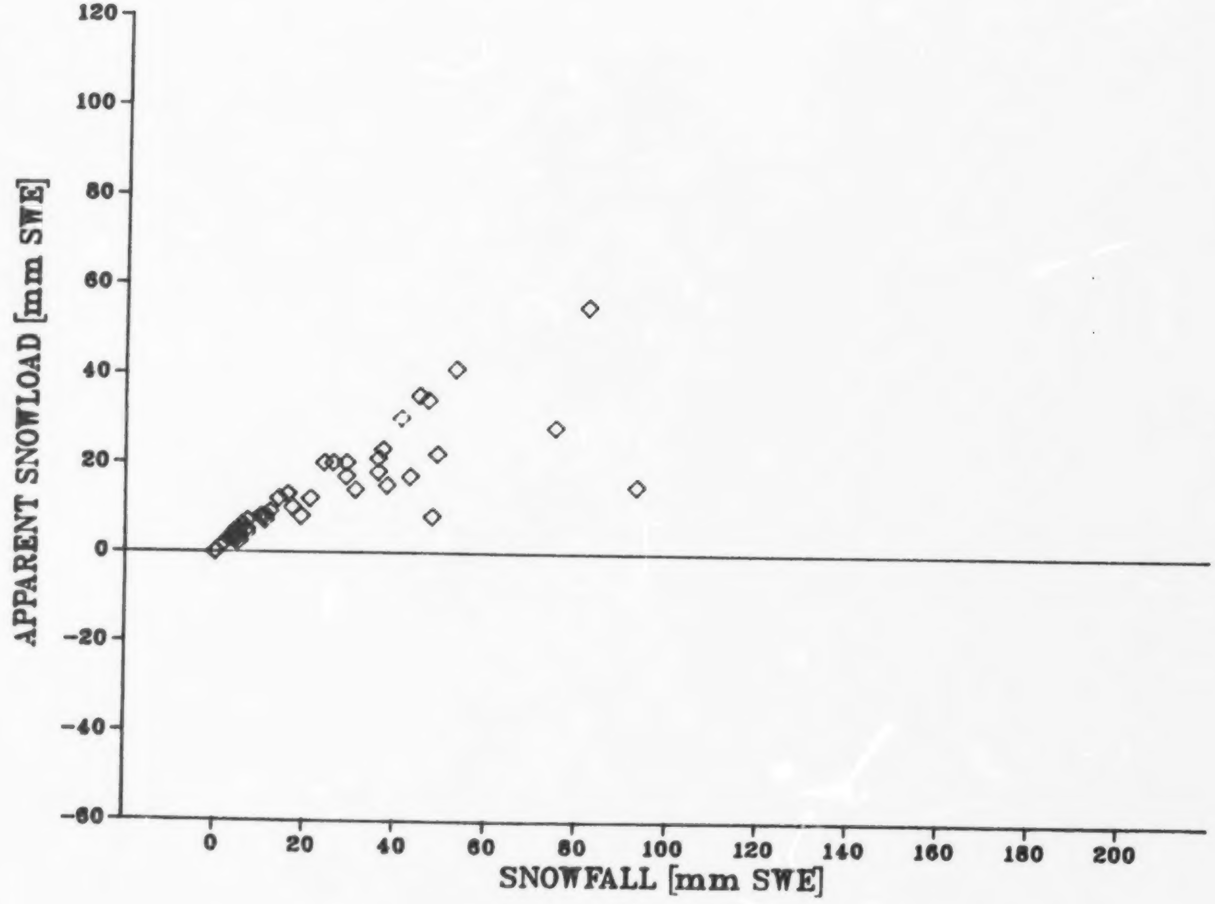

Figure 8.6 Apparent snow load in trees during individual storms at 710 m on Mt. Seymour (data of Fitzharris 1975: Appendix G).

same tree (e.g., Figs. 7.14 and 7.15). More interesting is the fact that there is no discernible sigmoid shape to the curves. Perhaps we should not expect the slow initial growth phase to be displayed due to the small amounts of SWE involved at that phase (<4mm) and the crudity of the technique (equating SWE in the open to snowfall, and equating interception to SWE in the open minus SWE under the canopy). More surprising is the lack of an upper asymptote. Only at the highest elevation (Fig. 8.1) do the data suggest an upper limit to the amount of snow a stand can intercept.

There are at least three potential reasons why the data for individual stands reveal no clear upper asymptote. First, the necessary crudity of techniques. Second, within stands the bridging between branches extends farther to bridging between branches of different trees. The upper asymptote is greater. Third, snowfall, particularly at lower elevations, was likely warm and wet. These latter conditions would increase the effectiveness of both adhesion and cohesion (Fig. 7.8) encouraging greater snow loads. However, such conditions held for the Japanese data on single trees as well (Figs. 7.14 and 7.15). It is noteworthy that apparent snow loads were sometimes negative (more snow was on the ground under the canopy than in the open). That condition most often prevailed at higher elevations where snowfall and apparent snow load was greater (Figs. 8.1 and 8.2). The observations suggest significant mass transport of snow from the canopy to the ground below the trees, a phenomenon discussed in

Ch. 7.2).

Fitzharris (1975) developed a regression equation describing the amount of snow under the canopy in terms of snowfall and elevation. He included the elevation term because both forest type and snowfall characteristics changed with elevation (e.g., Table 3.3). A "snowfall squared" term was included a priori to reflect the asymptotic approach to maximal snow load noted by JGFES (1952) and Satterlund and Haupt (1967) (see Figs. 7.13 and 7.14):

$$S(c) = -1.3 + 0.2\,S(o) + 0.0002\,S(o)\,H + 0.0013\,S(o)^2$$
$$(n = 511,\ r^2 = 0.78\ SE = 9.8) \qquad (8.1)$$

where S(c) = snow under the canopy (mm SWE), S(o) = snow in the open (mm SWE), and H = elevation (m).

The departure from linearity of Eq. 8.1 is slight (0.0013 $S(o)^2$). Given the large standard error of the regression, there seems little reason, other than purely theoretical ones, for including a power term. Note also that elevation itself plays a significant role in the summary equation (8.1) only at the highest elevations. The maximal elevation was only 1260 m, thus the 0.0002 H term cannot exceed 25% of snowfall in the open. The elevation coefficient is positive indicating more snow under the canopy (less interception) for equivalent snowfalls at higher elevations. Analysis of Fitzharris' data was repeated omitting data from elevations below 590 m where little snow persisted through the winter. The same r^2 and SE values were obtained. Next, the elevation term was dropped and a simple linear regression was computed. Equation 8.2 resulted, giving no change in the coefficient of determination and little change in the standard error.

$$S\ (c) = -4.5859 + 0.647\ S(o) \qquad (8.2)$$
$$(n = 380,\ r^2 = 0.78,\ SE = 9.9)$$

One effect of greater elevation was to increase snow on the ground under the canopy, or to reduce apparent interception (Figs. 8.1-8.6). Equation 8.2 indicates that the elevation effect was primarily through its contribution to S(o).

Data of Kittredge (1953) are not directly comparable to those of Fitzharris (1975). Whereas Fitzharris took all of his canopy snow measurements directly beneath the crowns of individual trees, Kittredge took his snow measurements under the canopy along transects in the forest. Fitzharris' data are uniform in their spatial orientation relative to individual trees; Kittredge's are not. Canopy cover therefore plays only a peripheral role in data of Fitzharris', but a primary one in data of Kittredge. The reverse is true for crown completeness.

Despite these differences we attain the same conclusion from data of both researchers. Within the data of Kittredge (1953) there is no asymptote for interception versus increasing precipitation which would relate to maximal storm size (Table 8.1). All of Kittredge's regression equations are linear.

Moving one level in the hierarchy of complexity, from interception processes in individual trees to interception processes in stands, seems to allow the emergence of new properties. Interception in stands does not appear to be the simple sum of interception by all trees in the stand. Whereas single trees appear to have a well-defined maximum snow load (Figs. 7.13-7.15), stands do not. Available data indicate that the amount of snow held in stand canopies increases linearly with increasing precipitation over a considerable range of precipitation. Data from Table 8.1 are presented in Figure 8.7. No tendency towards an upper asymptote to snow load or total interception is apparent (Fig. 8.7a).

As noted, a number of explanations for the absence of a clear asymptote are plausible:

1) Snow is blowing or falling off the canopy and being redeposited in the open. That redistribution would increase apparent snowfall, without increasing snow under the canopy. The fact that negative values of apparent snow load occurred most often at higher elevations where snow was drier (Figs. 8.1-8.6) and more likely to have been redistributed into the open, suggest that this phenomenon was not occurring frequently.

2) Significant amounts of snow are melting in the canopy and dropping off. This process would produce the negative

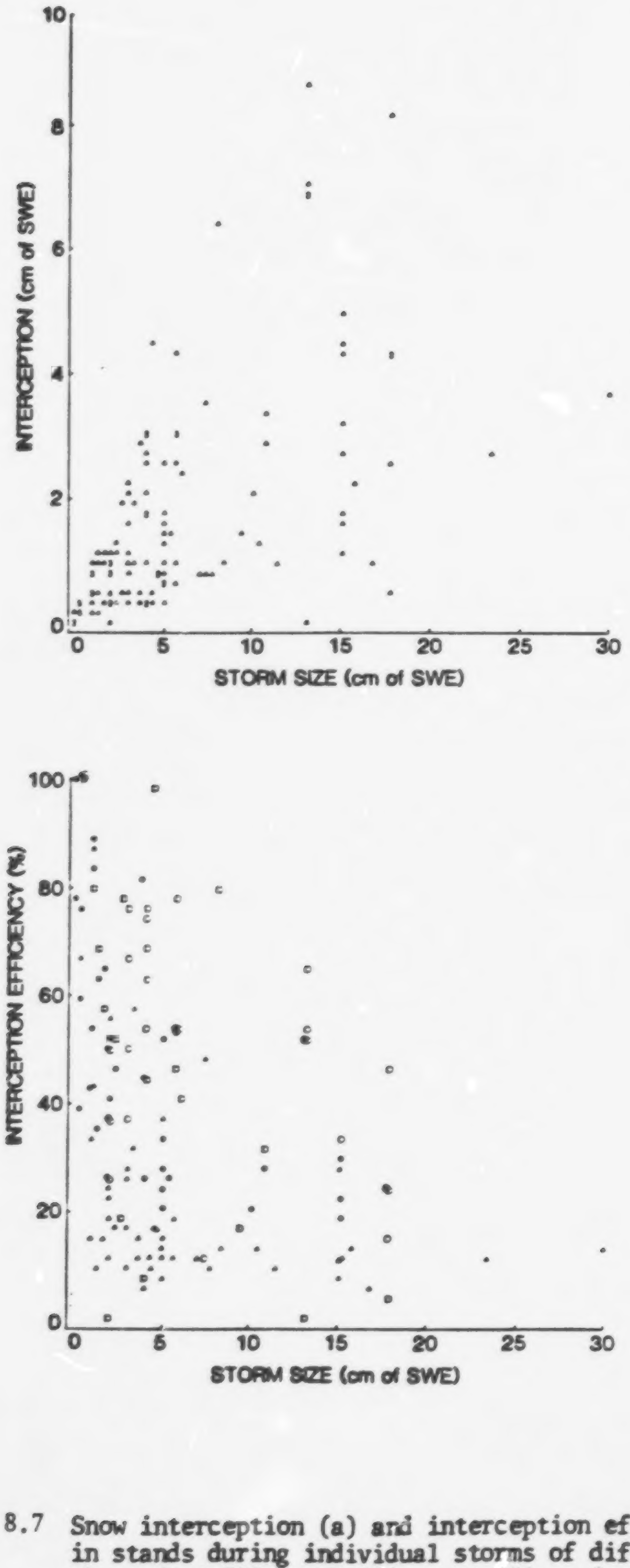

Figure 8.7 Snow interception (a) and interception efficiency (b) in stands during individual storms of different sizes where (○=0-30%, ▲=31-60%, ●=61-80%, and □=81-100%) are estimates of mean crown completeness (data of Table 8.1).

snow loads observed because snow under the canopy would increase relative to snow in the open.

3) Significant amounts of snow are sublimating or melting and being evaporated.

4) Adjacent trees are interacting in some manner (i.e., interlocking branches) so that greater snow loads can be held.

Although points 2) and 4) are likely true, we feel that the apparent difference between stands and individual trees is a somewhat spurious one. The premise that snow in the open equals true snowfall and that snow in the open minus snow under the canopy equals snow interception is unlikely to be wholly correct. If individual trees in a stand were each weighed during snow storms which were accurately measured by gauges positioned above the canopy, there seems little doubt that maximal snow loads could be measured for stands (e.g., Fig. 7.13). Kittredge, in his pioneering work, and many others after him simply assumed that intercepted snow sublimates rapidly. Actually we do not know how the proportion of total snowfall intercepted by a stand varies with total precipitation, but can derive first approximations. Broadly, interception efficiency decreases with increasing storm size. Data of Table 8.1 illustrate this broad pattern (Fig. 8.7b). Later (Figs. 8.9 and 8.10), we offer other evidence of the effects of storm size on interception by reanalyzing data of Kittredge (1953) using the approach of Harestad and Bunnell (1981).

8.1.2 Elevation.

Recognizing the potentially important effects of snow temperature on interception efficiency and maximal snow load (e.g., Chs. 7.2, 7.4), Fitzharris reanalyzed his data with respect to 3 functional elevation zones: i) the "drift snow zone" at 1260 m where snowfall was colder and drier, and redistribution by wind could render open versus canopy comparisons of negligible value in evaluating interception; ii) the "wet snow zone" located below the equivalent temperature where much precipitation fell as rain (Ch. 3.2); and iii) the "snow zone" located above the equivalent temperature but below 1260 m (Fig. 8.8).

The results were:

Drift Snow Zone

$$S(c) = 6.0 + 0.2\ S(o) + 0.0041\ S(o)^2 \qquad (8.3)$$

$(n = 82,\ r^2 = 0.73,\ SE = 15.2)$

Snow Zone

$$S(c) = -1.4 + 0.0006\ S(o)\ H \qquad (8.4)$$

$(n = 188,\ r^2 = 0.74,\ SE = 9.1)$

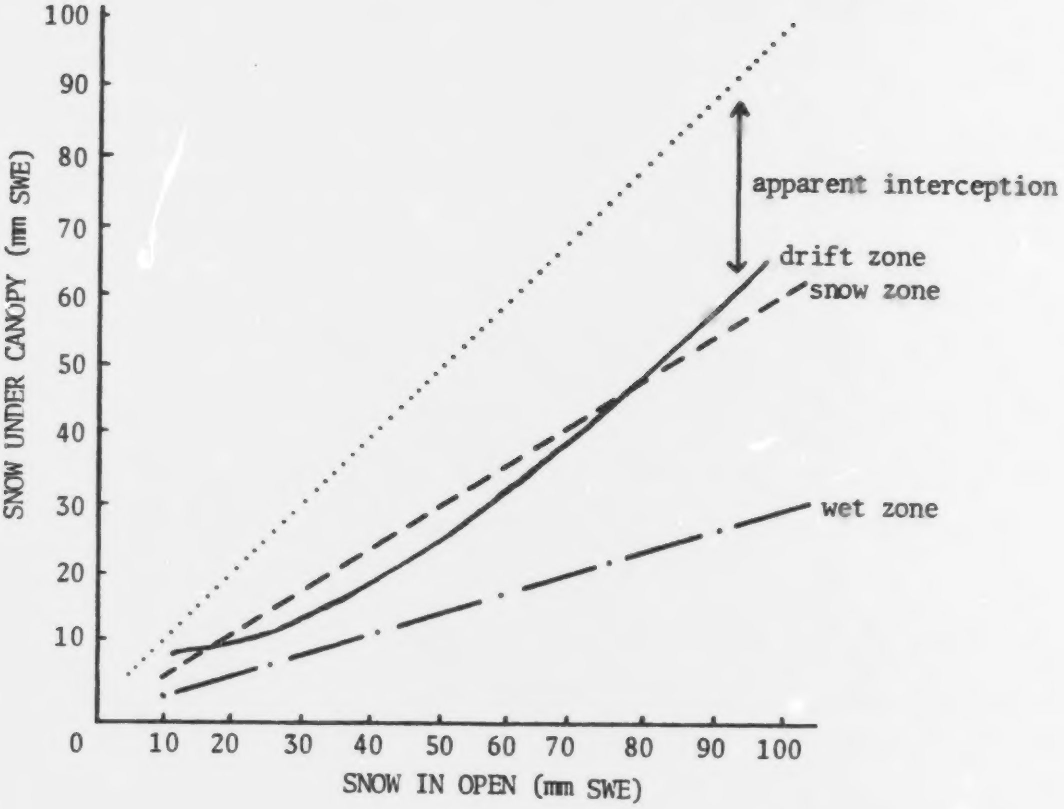

Figure 8.8 Regressions of snow under the canopy as a function of snow in the open for 82 storms on Mt. Seymour (data of Fitzharris 1975: 271).

Wet Snow Zone

$$S(c) = -0.7 + 0.0004\ S(o)\ H \tag{8.5}$$

$(n = 178,\ r^2 = 0.54,\ SE = 5.2)$

The large SE for the drift snow zone reflects the difficulty of the technique (snow in open minus snow under canopy) where redistribution of snow is a significant factor. However, the data imply more mass transport from moderate melting or shaking loose of heavy snow loads than redistribution to the open by wind. Orographic effects would produce higher wind speeds at higher elevations. The snow and wet snow zones exhibit equations for canopy versus open relationships which are linear (Eq. 8.4 and 8.5). Again, the relationships governing maximal snow load of individual trees seem to differ from those determining how much snow stands intercept in individual storms. No upper asymptote is evident. Note also that apparent interception was greatest in the wet snow zone where adhesive and cohesive forces were most effective (Fig. 8.8). The effects of temperature are more evident when data are stratified by elevation to approximate different temperature regimes. Slopes of regressions (Eqs. 8.4 and 8.5) are constant over storm size but differ with elevation class (Fig. 8.8).

8.2 Crown Closure and Efficiency of Interception

The difficulty in measuring the interceptive surface of a crown was noted earlier (Ch. 6, Ch. 7.6). Many researchers continue to utilize poorly understood "crown closure" or "canopy cover", often measured in an unspecified manner, as a predictor of snow interception. The analyses of Chapter 7.6 indicated that, for individual trees, some measure of vertical area was critical for accurate prediction of interception. Because heights are difficult to measure in a stand there are very few data explicitly addressing the role of crown height in interception by stands. Crown closure is the most common integrative measurement, and the one we examine. To reduce potentially confounding factors, analyses are here restricted to single storms when possible.

Figure 8.9 presents Kittredge's (1953) equations for SWE of intercepted snow transformed into percent interception. It is obvious that storm size has an overwhelming influence on percent interception causing within-stand ranges in interception of 10-100 percent. Interception efficiency decreases with increasing storm size, but decreases less rapidly when crown closure is greater (Fig. 8.9). The relationship with crown closure indicates that a unit of canopy is a relatively more efficient interceptor at higher snowfalls than at lower snowfalls. Figure 8.10 presents percent interception as a function of canopy cover for various storm sizes as derived from data of Kittredge (1953). With increasing storm size, the slope decreases. Within the data there is no apparent asymptote for snow load versus increasing storm size which would relate to maximal interception. The

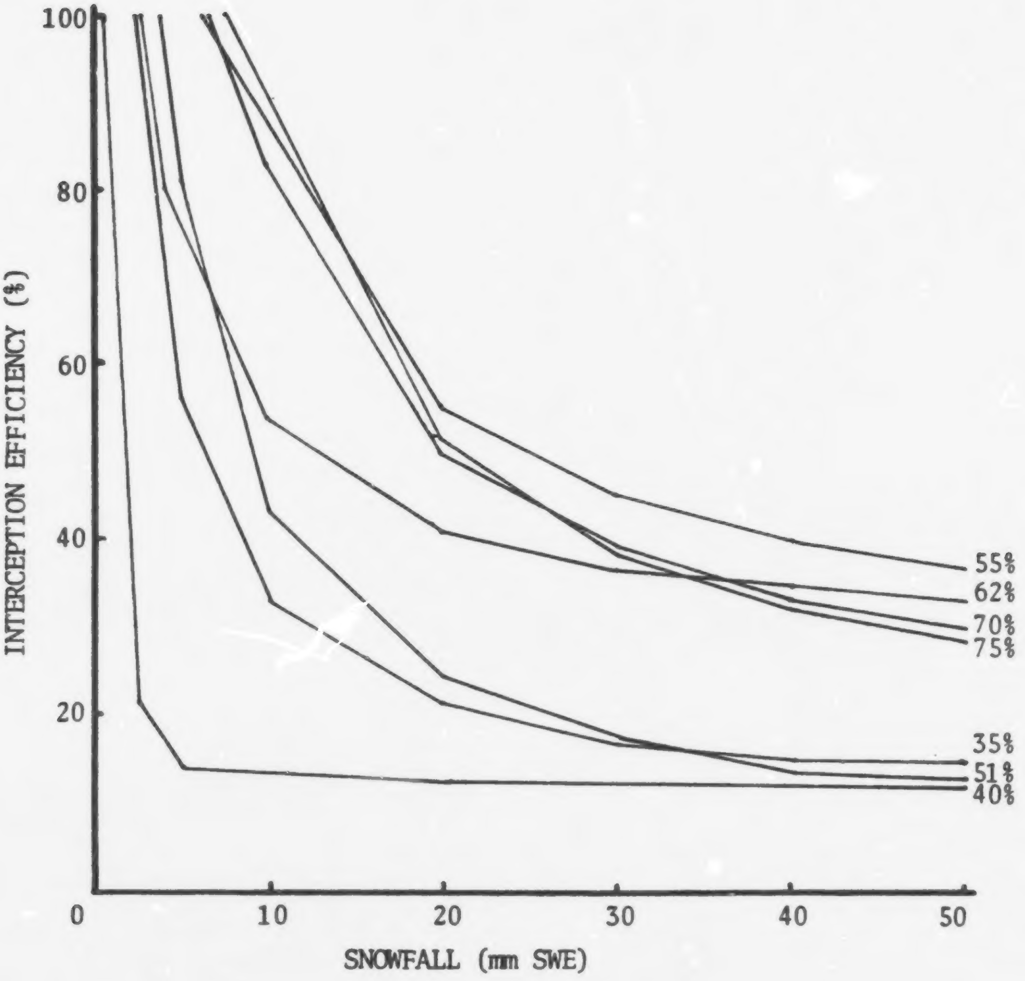

Figure 8.9 Effect of storm size on interception efficiency. Percentages are measurements of crown closure. (derived from equations of Kittredge 1953: 9).

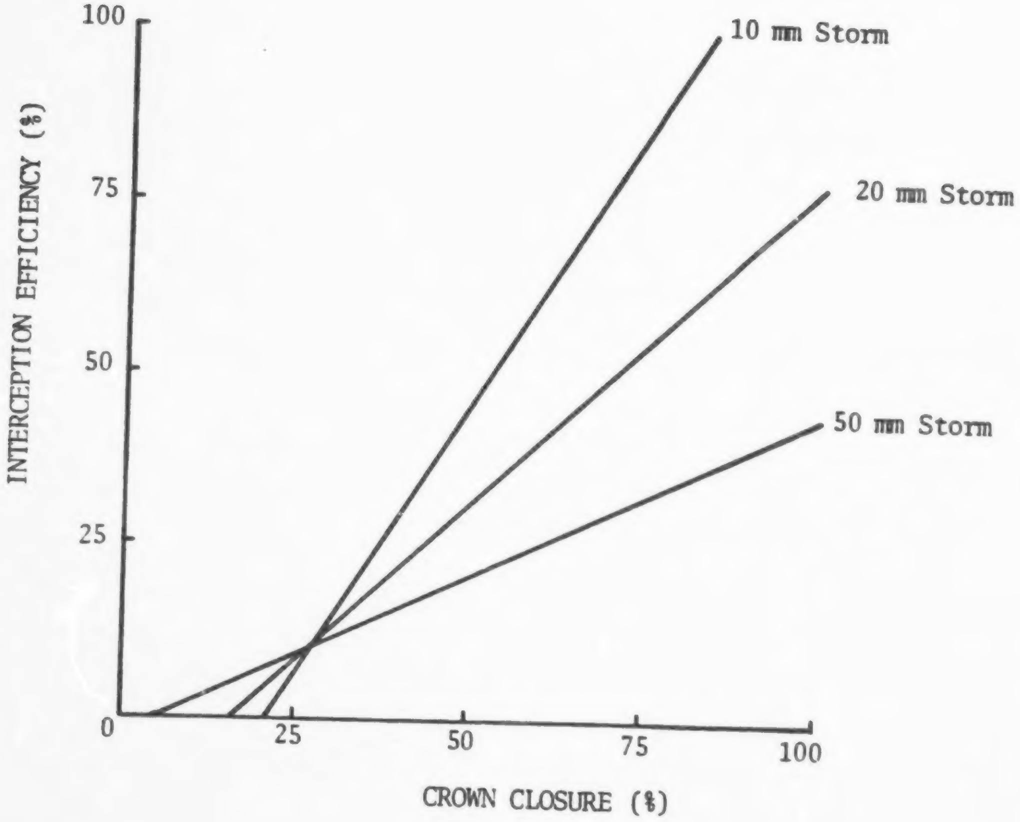

Figure 8.10 Effect of crown closure on percentage interception for various storm sizes (derived from data of Kittredge 1953: 9).

broad pattern is similar to that of interception efficiency by single trees (Fig. 7.18), and other stands (Fig. 8.7). All of Kittredge's regression equations are linear, but this may result ebcause the large variance precluded other interpretations of the data.

Data of Fitzharris (1975) illustrating the effects of storm size on interception efficiency (Fig. 8.11) contain more variability. His data were analyzed by broad elevation class. Crown closure differed between elevations; the "crown closure index", CCI, was 0.64 at 970 m (Eqs. 8.6 and 8.7) and 0.29 at 1060 m (Eqs. 8.8 and 8.9). Character of the snowfall also differed between elevations (Ch. 3.2).

Interception efficiency (%) at 970 m and CCI = 0.64

$$IE = 79.9 - 0.46\, S(o) \qquad (8.6)$$

$(n = 78,\ r^2 = 0.25,\ SE = 26.9,\ P < 0.001)$

Interception efficiency (%) at 1060 m and CCI = 0.29

$$IE = 78.07 - 0.63\, S(o) \qquad (8.7)$$

$(n = 73,\ r^2 = 0.25,\ SE = 28.5,\ P < 0.001)$

There is a clear influence of storm size on the amount of snow intercepted by the canopy (increasing with storm size, Eq. 8.3-8.5) and the interception efficiency (decreasing with increasing storm size, Fig. 8.11, Eq. 8.6-8.7). The broad pattern is thus similar to that observed in data of Kittredge.

Clearly, any attempt to predict percent interception by

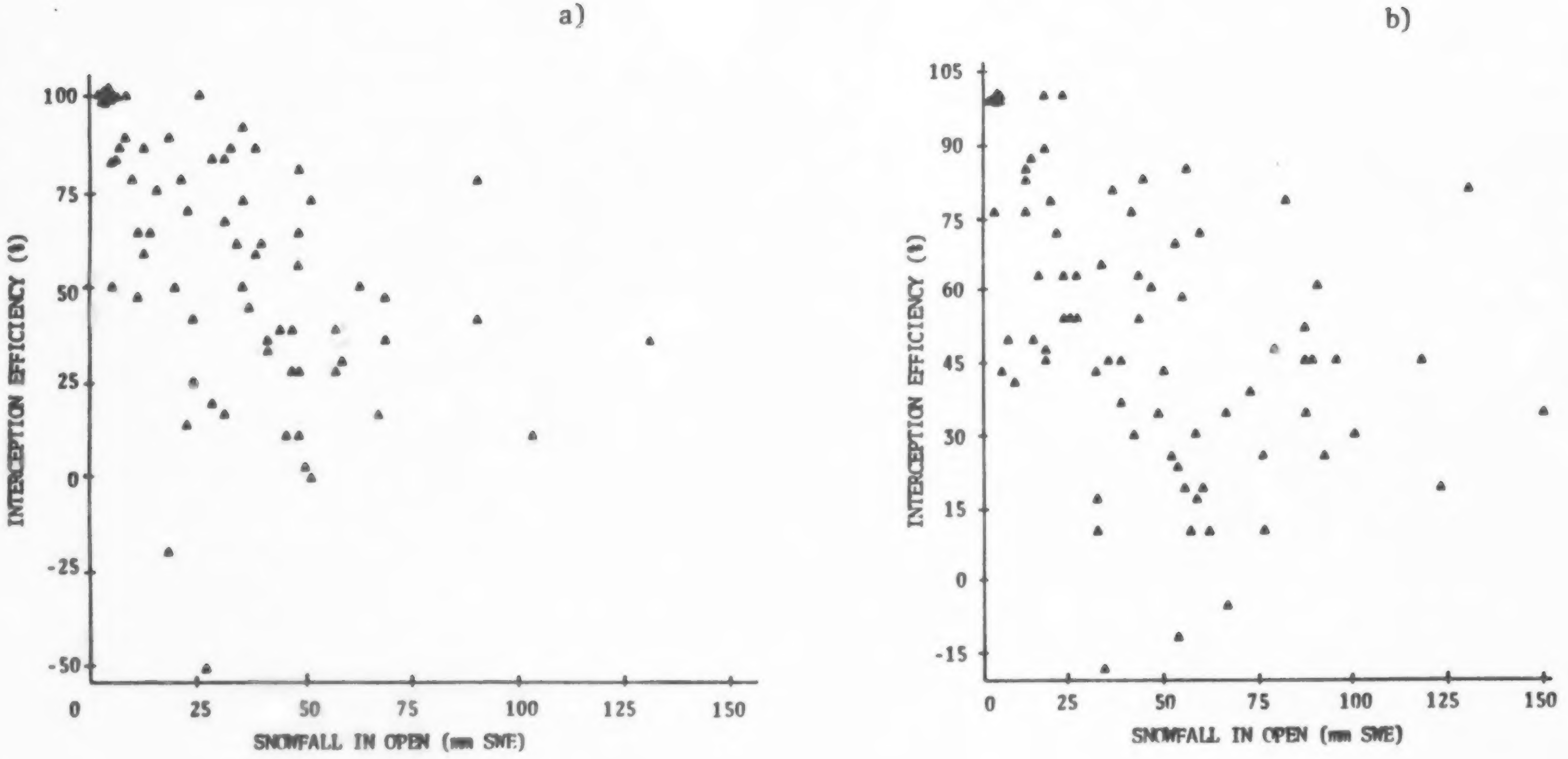

Figure 8.11 Effect of storm size on interception efficiency at 970 m (a), and 1060 m (b) elevation (reanalysis of Fitzharris 1975: Appendix G).

canopy measures alone is not appropriate; a storm size component must be included:

$$AIE = f(A,C) \tag{8.8}$$

where AIE = apparent interception efficiency (%), C = canopy cover (%), and A is some storm size function. Harestad and Bunnell (1981) suggested that A can be described by a function incorporating the slopes of regressions of relative SWE (SWE in forests/SWE in open x 100%) on canopy cover or crown closure for various snow regimes. The relative SWE is assumed to reflect a canopy's apparent interception efficiency (AIE) and allows inter-study treatment of data on a relative basis. It has the effect of normalizing snow regimes and allows the influence of canopy cover to be extracted more clearly. The relationship thus becomes:

$$AIE = 100 + A(C) \tag{8.9}$$

and $$A = a + b\,S(m) \tag{8.10}$$

where S(m) = maximum snow water equivalent in open.

The regression resulting from the data that Harestad and Bunnell (1981) presented is strongly linear:

$$A = -1.51 + 0.015\,S(m) \tag{8.11}$$

$(n = 13,\ r^2 = 0.82,\ SE = 0.19)$

Slopes of the regressions (Eq. 8.9) are strongly negative at low snow accumulations and weakly negative at higher snow accumulations. That is, interception efficiency is relatively less under conditions of deep snowfall.

Data of Fitzharris (1975) as well as data collected at the UBC Research Forest and more recent data from Mt. Seymour, were added to those collated by Harestad and Bunnell (Table 8.2). These data are from snowpacks. The maximum SWE is the maximum observed during the winter. Figure 8.12 shows that data outside the range presented by Harestad and Bunnell (1981) indicate a non-linear relationship. Fitzharris' data from high elevation areas (1060-1260 m) with large snow accumulations depart dramatically from Eq. 8.11.

In regions of low snow accumulation [low S(m)] it is reasonable to assume that snowfalls are infrequent with relatively little snow deposited. Interception efficiency is thus consistently high (Fig. 8.12). There are two possible explanations for the curvilinearity in regions of intense snowfall. First, at higher elevations, frequent intense storms are expected. Even though interception efficiency is lower during intense storms (Fig. 8.11), the greater frequency of large storms increases the total interception during the winter (Eqs. 8.3-8.7). Stated differently, a unit of canopy closure contributes more to apparent interception in regions of very high snowfall (the slope of the regression is steeper and relative interception is greater, Fig. 8.12). Recall, apparent interception is the difference between snow in the

Table 8.2 Effects of forest canopy cover on maximum snow-water equivalents (adapted from Bunnell and Harestad 1981).

Forest type	Stand age	Canopy closure (%)	Location	Elevation m	Slope of relative SWE canopy cover regression	Maximum SWE in open cm	Reference
Mixed hardwood and conifer	Saplings to sawtimber		New York	458-518	-0.30	27.2	Lull & Rushmore 1961
Lodgepole pine	-		Montana	High	-0.24	67.3	Farnes 1971
White pine	Various ages		Idaho	824-1678	-0.24	79.5	Packer 1962
Ponderosa pine	All ages		California	1525-1982	-3.12	4.1	Kittredge 1953
Ponderosa pine	All ages		California	1525-1982	-1.39	18.3	Kittredge 1953
White fir	140 years, mature		California	1525-1982	-0.70	50.8	Kittredge 1953
Red fir	200 years		California	1525-1982	-0.76	36.8	Kittredge 1953
Douglas-fir	Old growth & selectively logged		Oregon and Washington	503-534	-0.37	68.6	Kittredge 1953 (re-analysis of Hale 1950)
Mixed conifers	All ages		California	1525-1982	-1.20	12.7	Kittredge 1953
Mixed conifers	All ages		California	1525-1982	-0.93	38.1	Kittredge 1953
Mixed conifers	All ages		California	1525-1982	-1.05	53.1	Kittredge 1953
Mixed conifers	All ages		California	1525-1982	-1.33	24.9	Kittredge 1953
Mixed conifers	All ages		California	1525-1982	-0.52	83.0	Kittredge 1953
Mixed conifers	All ages		California	1525-1982	-1.43	7.1	Kittredge 1953
Mixed conifers	All ages		California	1525-1982	-1.08	21.1	Kittredge 1953

Table 8.2 (continued)

Western hemlock & yellow cedar	Mature	09	British Columbia	1260	-5.17	243	Fitzharris 1975
Western hemlock & yellow cedar	Mature	09	British Columbia	1260	-6.17	90	Fitzharris 1975
Western hemlock & yellow cedar	Mature	29	British Columbia	1060	-1.63	207	Fitzharris 1975
Western hemlock & yellow cedar	Mature	29	British Columbia	1060	-1.39	69	Fitzharris 1975
Western hemlock & yellow cedar	Mature	64	British Columbia	970	-0.36	130	Fitzharris 1975
Western hemlock & yellow cedar	Mature	64	British Columbia	970	-0.81	25	Fitzharris 1975
Western hemlock & yellow cedar	Mature	69	British Columbia	870	-0.65	78	Fitzharris 1975
Western hemlock & yellow cedar	Mature	71	British Columbia	790	-0.55	82	Fitzharris 1975
Western hemlock & yellow cedar	Mature	91	British Columbia	710	-0.48	60	Fitzharris 1975
Douglas-fir & western hemlock	60 years	97	British Columbia	525	-0.72	32.7	UBCRF data
Douglas-fir & western hemlock	80 years	91	British Columbia	740	-0.42	35.2	UBCRF data
Douglas-fir & western hemlock	120+ years	81	British Columbia	725	-0.36	41.2	UBCRF data
Douglas-fir & western hemlock	120+ years	97	British Columbia	580	-0.66	50	UBCRF data
Douglas-fir & western hemlock	120+ years	73	British Columbia	970	-0.75	59.7	Mt. Seymour data
Douglas-fir & western hemlock	80 years	82	British Columbia	970	-0.91	59.7	Mt. Seymour data

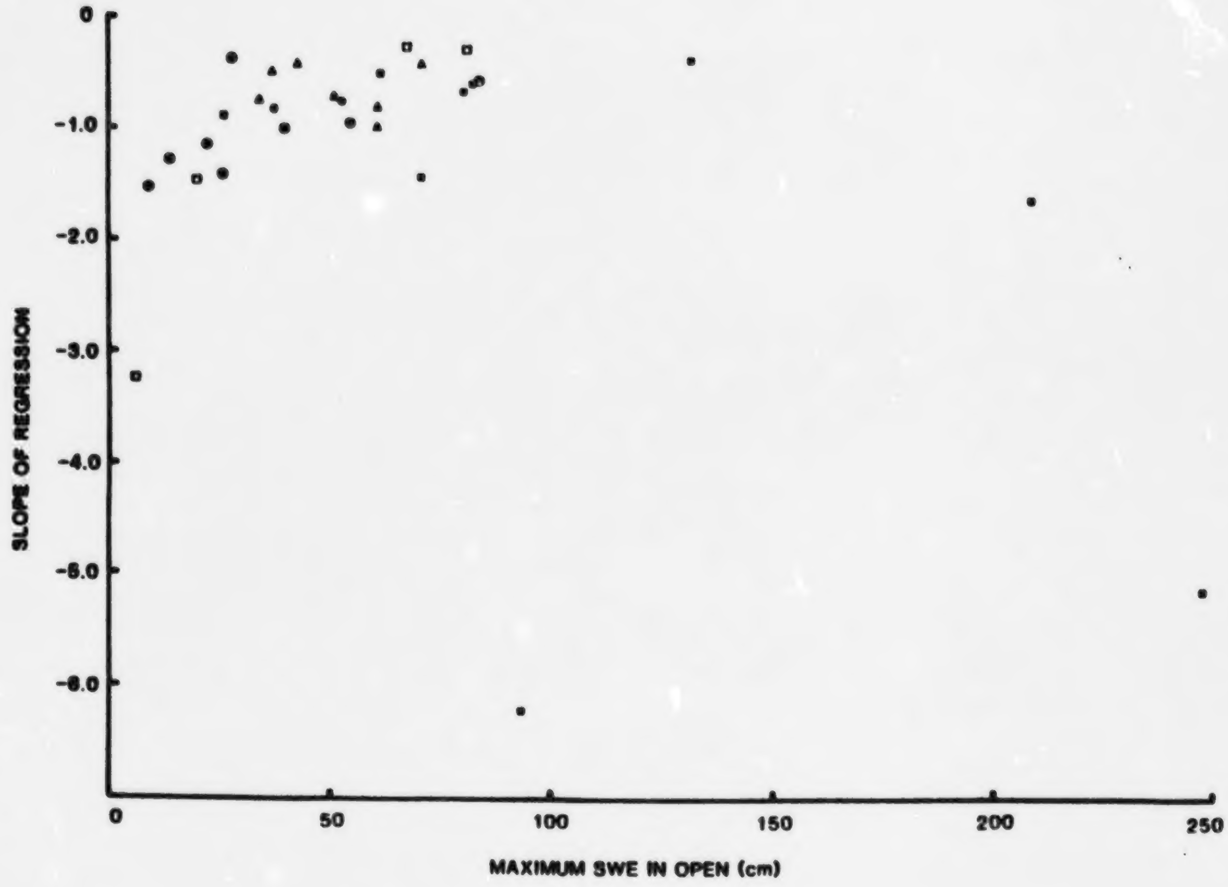

Figure 8.12 Slope of SWE/canopy cover regression as a function of maximum SWE in the open. SWE used in regression is the relative SWE (SWE in forest/SWE in open x 100). Symbols represent mixed hardwood (o), Mixed softwood (⊖), Douglas-fir and western hemlock (▲), western hemlock and yellow cedar (■), true fir (●), and pine (□) forest types.

open and under the canopy. Second, Fitzharris' high elevation data are from the dry snow zone where crown closure was only 9 to 29%. Any additional unit in crown closure is significant and the potential for mass transport into open areas is high, possibly biasing interception estimates upwards (Ch. 8.1.2).

The analysis here agrees with Harestad and Bunnell (1981) who noted that the relationship could prove curvilinear. The regression using the larger data base is:

$$A = -1.94 + 0.035\ S(m) - 0.0002\ S(m)^2 \qquad (8.12)$$

$$(n = 29,\ r^2 = 0.76,\ SE = 0.50,\ P < 0.001)$$

Omitting the two high elevation points for which no single, unequivocal explanation can be offered, yields a positive logarithmic function:

$$A = -3.1 + 0.62 \ln [S(m)] \qquad (8.13)$$

$$(n = 25,\ r^2 = 0.74,\ SE = 0.32,\ P < 0.001)$$

Apart from the change in shape of the relationship the general conclusion is still that canopy cover integrates age and species characteristics well, and can be used to predict SWE in coniferous forests. Because AIE and S(m) are SWE measurements from total snowpacks, the relationship must also integrate inter-storm ablation. When the approach of Harestad and Bunnell (1981) was applied to data for individual storms (Table 8.1), no significant relationship was apparent (P >

0.90).

Acceptability of the procedure developed by Harestad and Bunnell (1981) is dependent upon a proper understanding of the canopy cover measure employed. Figure 8.13 presents our data on the relationship between canopy cover and percentage interception in a single storm when canopy cover is measured by various means. Five of the seven measurement techniques employed were significant predictors of snow interception efficiency and are presented here. Neither the spherical densiometer nor the light meter seemed promising as predictors of interception efficiency. The data from Figure 8.13 are expressed as equations in Table 8.3. Data are aggregated for 8 different spacing designs [3 x 3 m (50%, 33%, and 0% thinned), 6 x 6 m, 9 x 9 m, 12 x 12 m, 15 x 15 m,and 3 Nelder plots] giving a total of 41 plots for each regression. All plots were within a juvenile (18-20 year old) Douglas-fir stand. Plot means were used in the regression analysis; within plots, n = 4 or 2 for each canopy measurement technique, and n = 32 or 16 for snow depth measurements.

Comparison of the statistics in Table 8.3 shows that the moosehorn (Bonnor 1967) has the highest r^2 value and lowest standard error. The moosehorn is followed in degree of predictive power by the ocular estimate, the photographic technique utilizing a subtended angle of 10°, and finally by the photographic techniques utilizing larger angles.

Wider angles used in canopy measurement incorporate more of the vegetation cover. At any point, measurements using

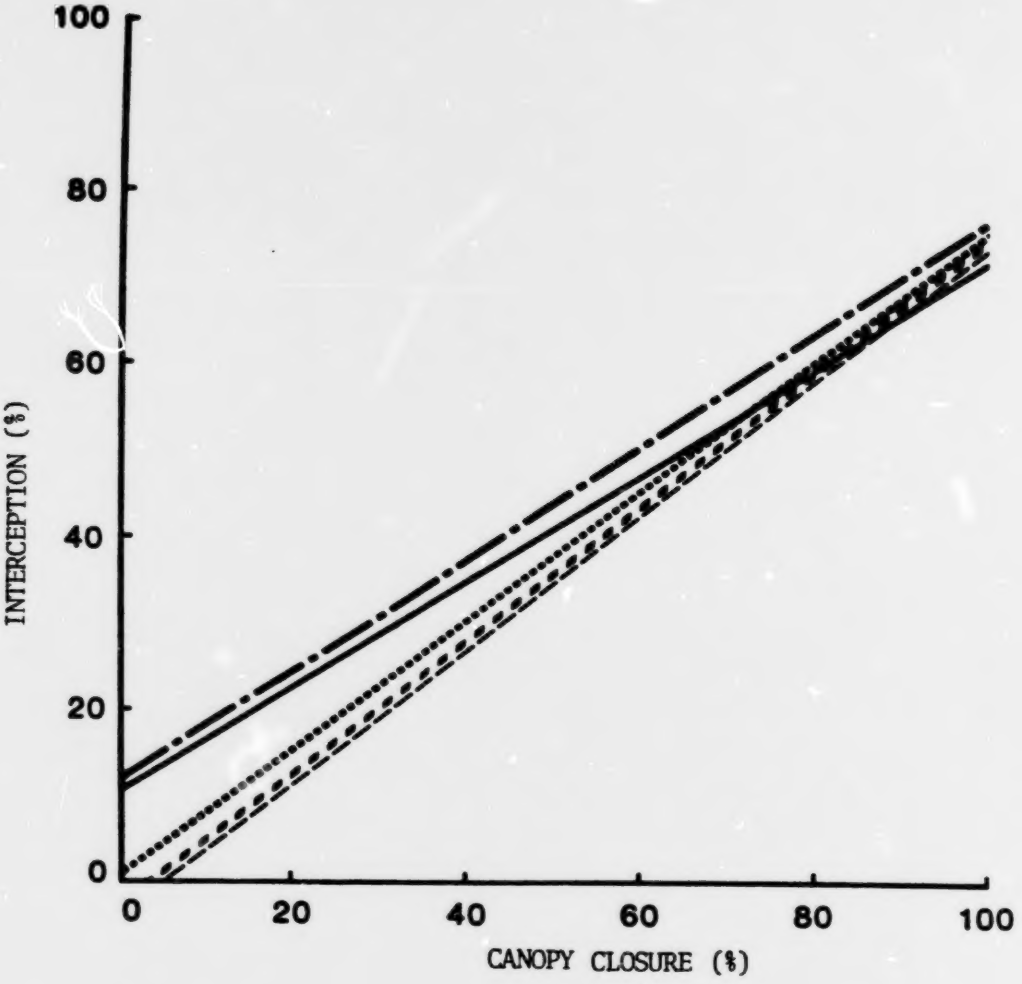

Figure 8.13 The effect of techniques of measuring canopy closure on regressions of interception versus canopy closure. Legend: (—·—) ocular estimate, (———) photo 10°, (········) photo 20°, (- - - -) photo 30°, and (— — —) moosehorn on regressions of interception and crown completeness.

Table 8.3 Regression equations relating percent interception (IE), during two single storms, to crown completeness (CC)[1] as a function of canopy measurement technique.

Canopy measurement technique	Equation	r^2	$S_{y.x}$	$P(\text{slope} \neq 0)$
Moosehorn	IE = 0.77(CC) - 4.39	0.74	8.88	<0.01
Occular estimate	IE = 0.65(CC) + 12.2	0.63	10.61	<0.01
Photo (10° cone)	IE = 0.62(CC) + 11.1	0.55	11.74	<0.01
(20° cone)	IE = 0.73(CC) + 1.52	0.53	12.02	<0.01
(30° cone)	IE = 0.78(CC) - 4.21	0.48	12.62	<0.01
Spherical densiometer		0.22		>0.01
Light meter		0.16		>0.01

1/ Measurements were taken over a range of canopy conditions in 8 different spacing regimes.

larger angles tend to yield higher canopy covers in young stands. As would be expected, higher y-intercepts and higher slopes necessarily result. This fact is evident in Table 8.3 when the statistics for the photographic techniques are compared. The angle subtended by the moosehorn is only slightly less than 10° and is the closest we have to a point measurement. It is unclear why it should provide the best estimator of interception efficiency. Ocular estimates were taken immediately after the moosehorn and may be biased. Except for the apparent anomaly of the moosehorn (which theoretically should yield the same values as 10° photos), there is a general tendency for the predictability of interception efficiency to increase with decreasing angle of measurement. The trend is expected in young canopies experiencing wet snow (there is little crown depth and the snowfall approximates vertical). The moosehorn may not prove as effective a predictor in old-growth where point measurements may be less important.

8.3 Forest Openings and Factors Influencing Snow Deposition

The effects of wind and temperature (radiation) on apparent interception occur simultaneously and are usually inseparable in existing data. The broad effects were noted in Chapter 6 and examined in Chapter 7. We noted that within continuous snow storms, snow loads on weighed trees varied as a function of both wind and temperature (Fig. 7.10). Abrupt declines in snow load during a snow storm were often associated with warming and large clumps of snow sliding from the branches. Increased wind speed also decreased snow loads (Fig. 7.10b). Both temperature and radiation are modified by openings in the forest, thus openings in the forest will modify apparent interception.

8.3.1 Radiation and Melting

Forest cover affects snow melt in three ways: i) the canopy very effectively reduces solar energy reaching the snowpack through shading; ii) the canopy absorbs solar energy and produces metabolic heat both of which are transmitted to the snow as long-wave back radiation; and iii) the canopy blocks long-wave heat loss from the snowpack to the clear, cold night sky.

To our knowledge, there has never been a comprehensive energy-balance study relating energy inputs and outputs to snow in forested conditions. Process-level data are few. Most data are of the "black box", non-causal nature which are useful primarily in a general, descriptive or comparative manner.

We can do little to extend our treatment of temperature (radiation) beyond that already provided (Ch. 7.2 and 7.4). Our review indicated that in coastal British Columbia, temperature may frequently modify apparent interception by encouraging mass transport of partly-melted snow to the area below the crown. Review of the JGFES (1952) data suggested that heat sufficient to melt 20% of the snow in the crown could cause significant mass transport. Data of Fizharris (1975) indicate that mass transport was significant even at higher, colder elevations (Figs. 8.1-8.6). As noted earlier there can be sufficient insolation at temperatures below freezing to generate melting and sliding of snow (Ch. 7.4).

Forest openings would permit more incident radiation on some crowns at the openings' edge thus encouraging mass transport. The snow sliding from the crown is of higher density than newly-fallen snow and probably is of importance on winter ranges primarily through its effects on shrub burial.

8.3.2 Wind and Redistribution

The effects of small openings on wind speeds are potentially more troublesome. Openings in the stand created to produce forage could be the recipient of wind-eroded snow from the canopy. In this section we begin examining that possibility by reviewing the general characteristics of air motion in stands. The treatment presents an extension of the general characteristics of wind introduced in Chapter 4.

Air motion in stands.--The study of man-made and natural wind breaks has produced an increased understanding of the air flow patterns caused by such barriers. The downwind effect of the barriers is proportional to the height of the barrier (Geiger 1961). Distances downwind are therefore, usually expressed in units of H where H is the height of the barrier.

Barry and Chorley (1976) reported that the denser the obstruction the greater its wind-breaking effect and thus the shelter created immediately behind it is greater. Figure 8.14 shows that while denser barriers create shelter immediately behind the barrier they also create greater turbulence. The result is that the overall disruption in the flow of air terminates within a relatively short distance downwind from the barrier (Barry and Chorley 1976). According to Barry and Chorley, the maximum protection is given by barriers that have approximately 40% penetrability. In terms of a forest barrier the percent penetrability would depend on factors such as the presence or absence of ground vegetation, the depth of crown, and the species-specific characteristics of the crown-foliage area. Shiotani and Arai (1954) measured crown density and mean height to canopy as an index of penetrability; Sturrock (1972) used only crown density (Table 8.4).

Barriers placed in open terrain begin to influence air flow about 18 H upwind; the downwind effect lasts 20-25 H (Geiger 1961). Open forest stands act as a barrier and destroy wind vorticies both in the forest and in the adjacent areas up to 30 H distance to the leeward (Anderson 1970). Data of Table 8.4 show that for forested barriers the influences on wind flow is variable depending on the vegetation type, width, height, and density. The downwind

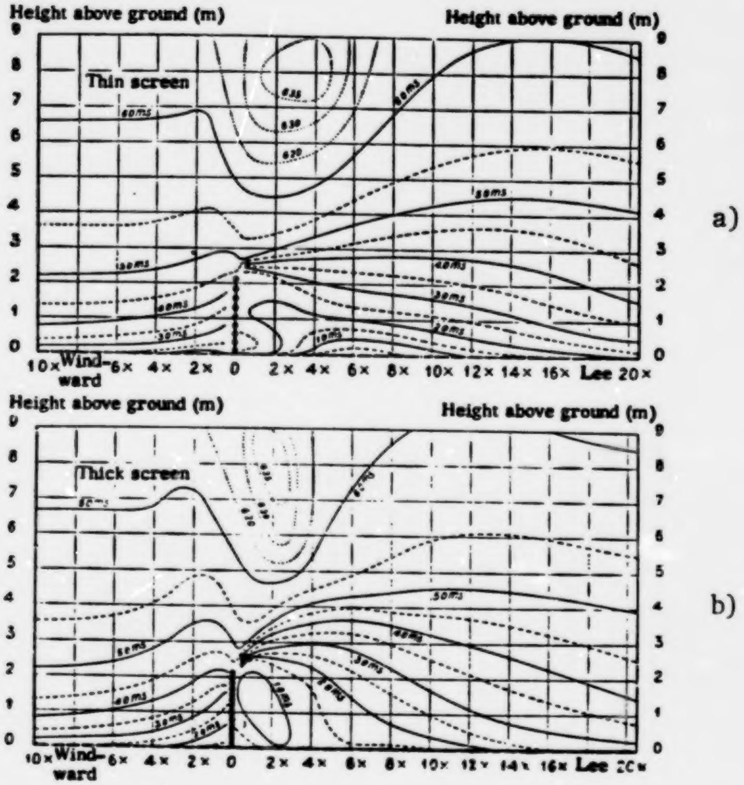

Figure 8.14 Wind field around two reed screens of different density (from Geiger 1961: 501 after Nägeli 1953: 231).
a) 45-55% penetrability.
b) 15-20% penetrability.

Table 8.4 Distances over which different barriers affect the pattern of wind flow.

Source	Height of barrier (m)	Width (m)	Penetrability	Type of barrier	Distance of effect Downwind	Distance of effect Upwind	Comment
Gary (1975)	10.3	-	-	Pinus contorta	5H[1]	-	Barrier was the windward forest, leeward forest influence on air flow is suspected
Finney (1934)	1.25	-	50%	Snow fence	16-20H	-	
Barry and Chorley (1976)	2.2	-	40%	Fence	20-25H	-	
Nägeli (1946)	1.4	-	low	Fence	27H	-	
			medium	Fence	30H	-	
			high	Fence	24H	-	
Shiotani and Arai (1954)	3 5	30 34	0.54 0.63	Cryptomeria japonica	15H >>10H	5H 10H	Penetrability given as density of crown layer
Sturrock (1972)	17.7	8.8	50%	Pinus radiata	18H	2H	
	26.0	11.6	50%	Pinus radiata	13H	1H	
	12.2	8.5	50%	Pinus radiata	17H	2H	
	12.2	12.8	50%	P. radiata and Cedrus deodora	22H	3H	
	8.5	5.1	50%	P. radiata and C. deodora	17H	2.5H	
	6.1	7.3	50%	C. deodora	16H	2H	
	13.1	6.4	50%	Cupressus macrocarpa	20H	4H	
	12.2	26.0	50%	C. macrocarpa	21H	1H	
	19.8	10.7	50%	Eucalyptus globulus	13H	1H	
	11.3	4.0	50%	Populus nigra	9H	0	Leafless
	11.3	4.0	50%	P. nigra	12H	0	Half expanded
	11.6	4.0	50%	P. nigra	12H	0	Fully expanded

1/ H = height of barrier

effect can be increased by back-coupling of more than one barrier (Fig. 8.15).

Summarizing the data of Table 8.4 reveals the following conclusions:

1) Increasing the height of wind barriers increases the distance downwind that air flow is affected. Upwind distance of the effect is also increased.

2) Increasing the width of wind barriers increases the distance downwind that air flow is affected. The upwind effect is reduced.

3) There appears to be a curvilinear relationship between penetrability and the downwind effect on air flow. For forest windbreaks, a penetrability of 40% gives the maximum distance of downwind effect.

Figure 8.16 shows the pattern of air flow over a snow fence of 50% density. A strong eddy effect lasts for up to 16-17 H downwind and a complete reversal of wind direction is brought about within that zone. Similar patterns were observed and measured by Bergen (1976) and Gary (1975) in their studies concerning air flow in small forest openings. Bergen found the recirculatory eddy to be centred on the lee edge and at the level of maximum foliage concentration (Fig. 8.17). The effect of the clearing lasted for more than

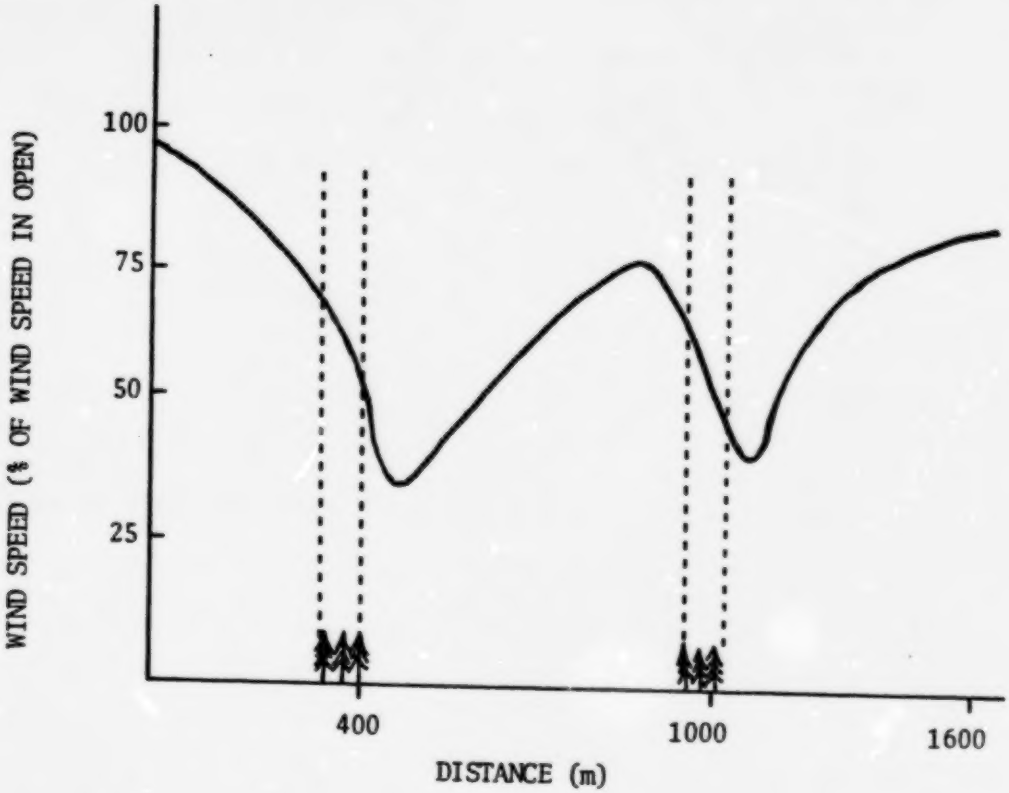

Figure 8.15 The effect of shelterbelts on leeward windspeed (redrawn from Geiger 1961: 503).

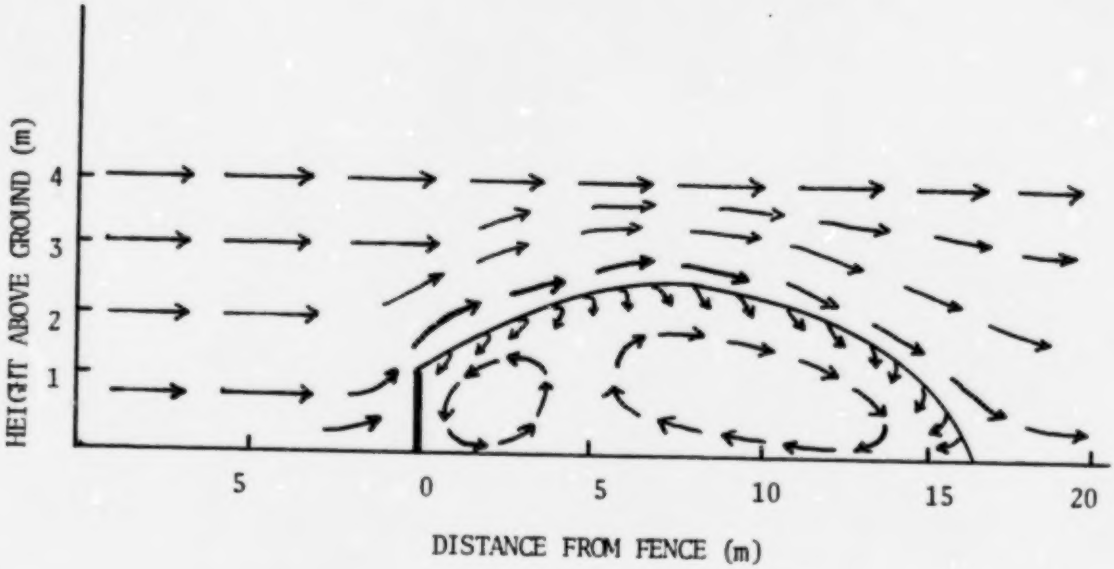
HEIGHT ABOVE GROUND (m)
4
3
2
1
5
0
5
10
15
20
DISTANCE FROM FENCE (m)

Figure 8.16 Pattern of air flow over a snow fence of 50% density (redrawn from Gary 1975: 111 after Finney 1934).

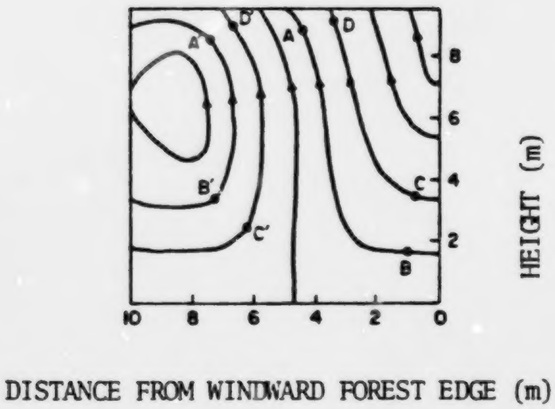

Figure 8.17 Estimated streamline pattern for windflow across a forest clearing (from Bergen 1976: 124).

5 H downwind (Bergen 1976); the eddy existed from 1/2 H upwind to 1/3 H downwind of the clearing's lee edge. While the most obvious feature of Figure 8.17 is the recirculation zone centered about 7 m high and about 2 m ahead of the lee clearing edge, there is a dividing streamline intersecting the clearing floor at the middle of the clearing. Most of the clearing is occupied by level or ascending air; the compensating subsidence is apparently confined to the stand behind the clearing (Bergen 1976).

Velocity of flow.--Kittredge (1948) found that a barrier's ability to reduce wind velocity increased as the initial velocity of wind increased. The relative reduction did not increase as dramatically. Gary (1975) measured the surface roughness and the wind profile in an undisturbed forest stand and foliage weight at selected intervals in height. Comparing the two variables reveals the inverse relationship between foliage weight and the ratio of local wind speed at a specific height to friction velocity at the ground surface (Fig. 8.18; see also Eq. 4.3). Wind speed is lowest at the ground increasing to a subcanopy maximum and then decreasing to a secondary minimum where crown foliage is most abundant. The variability in wind velocity reflects the variability in penetrability.

The subcanopy wind flow varies with the vegetation structure at that level. Penetrability may be reduced only slightly by tree stems. A dense layer of undergrowth,

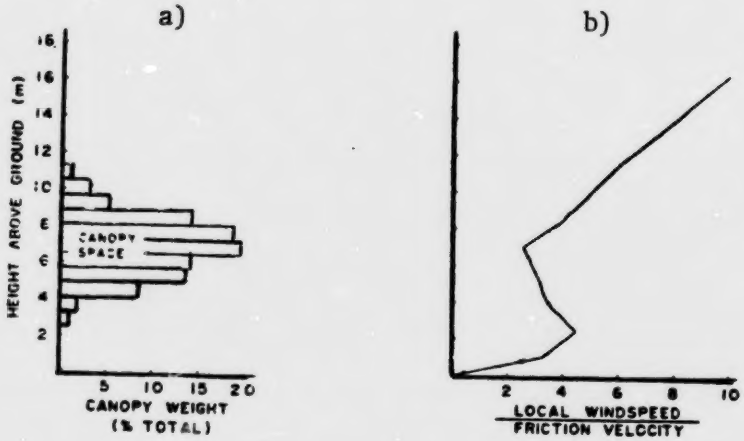

Figure 8.18 Wind profile as a function of canopy profile (from Gary 1975: 109).
a) Relative canopy weight distribution of an unbroken forest stand.
b) The average wind speed profile.

however, may eliminate the subcanopy maximum and reduce wind velocity to 20% of that above the canopy (Barry and Chorley 1976). Table 8.5 allows comparison of wind velocity in forest clearings as a percentage of the wind velocity above the windward forest barrier. From the data summarized we can make two tentative conclusions:

1) Wind speeds in clearings adjacent to uncut forests are much reduced from those on the windward side of the barrier (generally 50 to 95% less).

2) As the size of the opening decreases the reduction in wind speed in the opening increases.

Erosion by Wind.--Adhesive force between a snow particle and the intercepting surface is influenced by both wind speed and snow type (Figs. 7.11 and 8.19). As soon as snow particles settle, the cohesion between them begins to increase (Figs. 7.20 and 7.21). Unless wind acts concurrently with snowfall or directly afterward, its erosion potential on interception is greatly reduced. Erosion and mass throughfall during a storm were noted earlier (Fig. 7.10). In some areas, erosion of snow after a storm and its subsequent transport appear to be important. Such movement of snow would be affected by opening size and the influence of opening size on wind.

Hoover (1960) considered wind to be the causal factor

Table 8.5 Influence of tree density and opening size on wind speed.

Authority	Penetrability (% cut)[a]	Size of opening (H)[b]	Wind speed (m/sec)	% of wind in open[c]
Anderson et al. (1976)	100		0.60	100.0
	50		0.40	66.7
	0		0.10	16.7
Bates (1924)	100		1.02	100.0
	75		0.62	63.0
	65		0.89	89.0
	0		0.40	40.0
Haupt (1973)		0.50	1.35	83
			1.70	76
		2.25	1.36	84
			1.98	89
Jemison (1934)	100		0.62	100.0
	50		0.36	58.0
	0		0.09	12.0
Swanson (1980)		20		48.2
		6		9.3
		4		7.2
		2		4.4
		1		2.1
		0		3.0
Bergen (1976)		1		100.0
Brown (1972)		3		45.0
Sturrock (1972)		20		80.0
		10		50.0

a/ % cut refers to the number of stems/ha

b/ H = tree height of adjacent stand

c/ Measured as % of that over the windward barrier

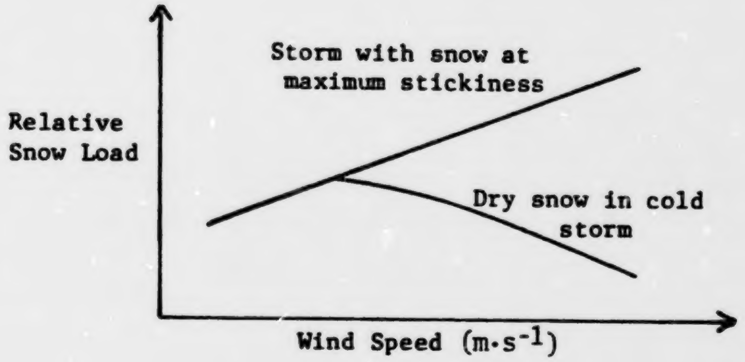

Figure 8.19 Schematic representation of the maintenance of loads of dry and sticky snows by trees as a function of wind velocities (from Bunnell 1978: 27).

creating discontinuous snowpacks in the alpine areas of the southern Rocky Mountains. Wind erodes snow from exposed areas and deposits it in sheltered areas.

Deposition of eroded snow is a much different consideration from that of new or freshly-fallen snow. Given time, the particles characteristic of fresh snow become enlarged and solidify into a more dense and more closely packed arrangement. Not only do the snow particles undergo considerable change before becoming airborne, but they undergo considerable change while aloft (Miller 1962). Those changes may range from reduction of particle size and mass to complete sublimation of a particle before it is redeposited (Tabler and Schmidt 1973).

Calculations of transport distances of eroded snow have been made only for areas of dry snow. The potential distance transported can be calculated given that air temperature, relative humidity, solar radiation, particle speed, and particle size are known. It is assumed that particle speed equals wind speed 2 m above the ground and that particle density is that of ice ($0.916\ \mathrm{g \cdot cm^{-3}}$). The calculations then provide an estimate of the distance a particle would travel before complete sublimation occurs providing gravity does not influence how long the particle is suspended in the airstream. Tabler and Schmidt (1973) developed the relationship used (Eq. 8.14).

$$Rm = \frac{0.715\ (U\ -\ 5.87)\ X^3}{(dm/dt)\ X} \qquad (8.14)$$

where Rm is the transport distance for an average-sized snow particle before it completely sublimates, and U is the drifting speed of snow particles at a height of 2 meters above the ground ($m \cdot s^{-1}$). U is derived from calculating the average horizontal velocity of a vertical column of snow (Up), given the wind profile and mass of drifting snow. Up is then expressed in terms of the wind speed at 2 meters thereby creating U . X is the average snow particle size and (dm/dt)X is the particle sublimation rate where:

$$\frac{dm}{dt}X = \frac{0.000895\ X\ (1+37X\ )(1-RH/100) + 0.000000522\ (85.6-T)X\ ^2S}{T^2 - 8.2T + 254} \qquad (8.15)$$

where RH is the relative humidity, T is the temperature in °C, S is the solar radiation in $cal \cdot cm^{-2} \cdot h^{-1}$ and X is the average sublimation particle diameter (cm). Sublimation diameter is assumed to be 1.1 times the mean particle diameter.

From equations 8.14 and 8.15 the dry conditions of Wyoming would allow particles of 0.015 cm diameter (the mean) to travel 900 m before complete sublimation (Tabler and Schmidt 1973). Equation 8.14 indicates that wind speed must exceed 5.87 $m \cdot s^{-1}$ (21.13 $km \cdot h^{-1}$) before transport of snow particles (with a density equal to that of ice) can occur. Data of the Atmospheric Environment Service from Vancouver Airport yield a mean wind speed during winter months of 2.62 $m \cdot s^{-1}$ (9.43 $km \cdot h^{-1}$). These speeds could be greater at higher elevations, but would be reduced by forest canopies. Other than gusting we should expect little transport of snow eroded from well-settled snow surfaces. To obtain some value comparable to that of Tabler and Schmidt (1973) we have used a wind speed of 7.0 $m \cdot s^{-1}$ (25.2 $km \cdot h^{-1}$) in Equation 8.14. The other values used (Eq. 8.15) are means extracted from data of the Atmospheric Environment Service or Chapter 3. They are:

Mean particle size	0.5 cm (aggregate; Fig. 3.27) 0.05 cm (sector plate; Fig. 3.23)
relative humidity	84%
temperature	0.0°C
solar flux	112.3 $cal \cdot cm^{-2} \cdot h^{-1}$

Using Equations 8.14 and 8.15, the distance that a sector plate would travel in coastal British Columbia before complete sublimation is 776.5 m; an aggregate of mean diameter could potentially travel 9,582 m. The longer transport distances calculated for coastal British Columbia compared to Wyoming, are a function of larger particle sizes and higher relative humidity. Most other variables (higher temperature, greater solar flux, and lower wind speed) should produce a shorter transport distance.

Tabler and Schmidt (1973) noted that particle speed as a function of wind speed still needs considerable research before confidence can be placed on the ditance that wind can transport snow. Direct extrapolation of their equations to coastal conditions, suggests that some particles could be transported considerable distances at wind speeds of 7 km·h^{-1}. What would actually happen is equivocal. Tabler and Schmidt derived their equations for particles much smaller than commonly occur on the coast and assumed a density of ice. The interactions of larger mass with gravity or increased drag are unclear, but would certainly reduce transport distances on the coast. Similarly, the most frequently occurring particles in coastal storms are needles and lump graupel (Table 3.5). Only graupel would approximate the density of ice (Ch. 3.3.2); sublimation rates would be greater from other particles. Transport of needles would more closely approximate the distances calculated by Tabler and Schmidt (1973) for Wyoming; graupel is sufficiently massive and has such high terminal velocities (Fig. 4.8) that transport distances would likely be much reduced. The moister coastal conditions would also strengthen adhesive and cohesive forces. The wind speed necessary to produce snow erosion on the coast is likely greater than the value of 21.1 km·h^{-1} assumed by Tabler and Schmidt (1973). Mean wind speeds during winter on the coast were 9.4 km·h^{-1}.

We conclude that application of available relationships cannot predict transport of eroded snow under coastal conditions. Given the apparent frequencies of different particle types on the coast it seems likely that transport is seldom great other than during wind velocities exceeding 25 $km \cdot h^{-1}$.

Summary.--Snow accumulation and pattern of deposition in clearings is a function of: i) the opening structure; ii) the velocity and pattern of wind flow there as influenced by both the windward and leeward forests; and iii) the characteristics of snow deposited during a storm.

8.4 Observed Patterns of Accumulation

In Sections 8.1 through 8.3 we have identified and discussed the underlying processes influencing interception in stands. Published research revealing information on these processes are rare because most research involves snowpack measurements as opposed to individual storm measurements and observations. Although snowpack measurements are frequent in the literature, they are confounded by the many processes interacting and occuring concurrently (Chs. 1 and 7). Additionally, most research concerning observed patterns of accumulation has utilized large openings as control areas. Snow in the forest is normally compared to snow in the open to reveal information concerning interception. Miller (1966) pointed out that this procedure is inappropriate. Reasons for our lack of confidence in this procedure are stated in Section 8.1. The difference between snow in the open and snow in the forest (our 'apparent interception', Miller's "ΔA") fluctuates with exposure of the forest stand and the floor of the openings to winds which change during the progress of a storm. Therefore, ΔA is increased by biasing the open catch upward. The value of ΔA also fluctuates with factors governing deposition (Chs. 4 and 7) and with factors governing ablation (Ch. 8.3.1).

In the following section we present published observations on the patterns of accumulation as influenced by forest stands. Because much of the data are from snowpack measurements, the results seem variable and sometimes contradictory. It is evident, however, that elevation and the size of forest openings are two major factors governing the influence on snowpack accumulation in forest stands.

Forest openings have their greatest effect in areas where dry snow may be blown about or redistributed by wind. We would expect no effect if snowfall was vertical; that is, during calm conditions or when snow was too dense to be moved by prevailing winds. When winds are present the windward side of an opening may experience an eddy effect (Fig. 8.20) with enhanced snow deposition whereas the leeward side may show lower accumulations due to higher wind speeds gathered across the opening. As we discuss in the following section, snowpack also increases in depth with increasing elevation due to greater snowfalls and slower rates of melt. In practice, it is difficult to separate these factors in data available which consist primarily of occasional measurements of the snowpack - often on or near April 1. We describe the effects of

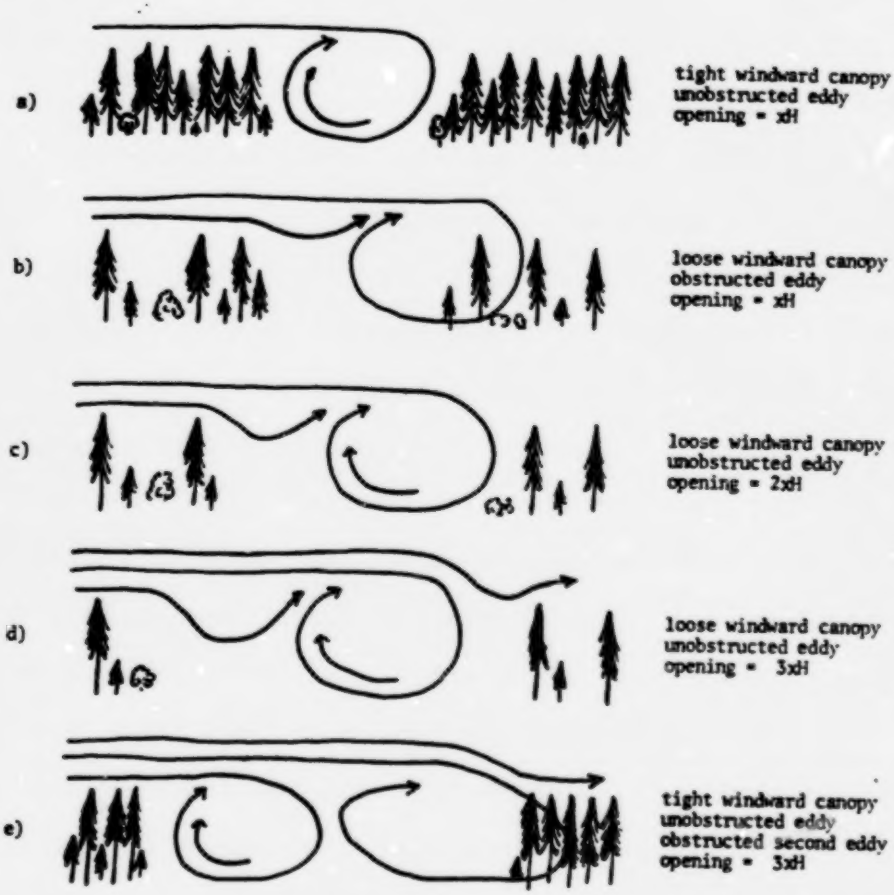
a)
tight windward canopy
unobstructed eddy
opening = xH
b)
loose windward canopy
obstructed eddy
opening = xH
c)
loose windward canopy
unobstructed eddy
opening = 2xH
d)
loose windward canopy
unobstructed eddy
opening = 3xH
e)
tight windward canopy
unobstructed eddy
obstructed second eddy
opening = 3xH

Figure 8.20 Schematic representation of eddy formation in forest clearings of different sizes and with different penetrability characteristics of the adjacent forest.

elevation on snowpack without reference of the relative weights of the fators creating this influence.

8.4.1 Elevation Effects

In a given storm the snowfall apparently increases almost linearly with elevation, although orographic effects may introduce modest curvilinearity at higher elevations (Ch. 3.2). The water equivalent of the snowpack, however, increases with elevation in either a curvilinear or linear manner depending upon the area. Schaerer (1970), in his analysis of B.C. snowpacks, suggested that the increase was curvilinear in a wet, mild climate and approximately linear in a dry, cold climate. Figure 8.21 presents Schaerer's data from several places in British Columbia illustrating the change in the form of the relationship. The curvilinearity comes from a change in rate of snow deposition with change in elevation. Melt at low elevations may occur simultaneously with snow deposition at higher elevations. Variation in the rain/snow boundary elevation exists from storm to storm and between years (Ch. 3.2). Generally, however, a greater number of snow-depositing storms with more snow causes a higher rate of deposition at higher elevation. Drier and colder climates, such as by Lake Louise (Fig. 8.21) have less potential for mixed rain and snow events than do coastal areas like Mt. Seymour. As a result, snowpack depth increases linearly with increasing elevation in colder, drier areas.

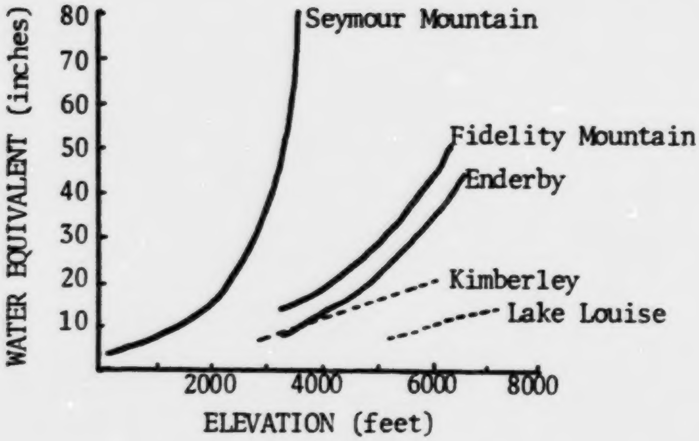

Figure 8.21 The effect of climate; cold and dry (-----), warm and wet (——), on relationships between snow accumulation and elevation (from Schaerer 1970: 56).

Claus (1981) extended Shaerer's analysis using National Research Council data from 18 mountains extending in a west-east transect from Vancouver to Fernie. His regressions used the means (over 3 to 12 years) for 6 to 15 stations at different elevations on each mountain. For mountains in southwestern British Columbia, he had data for 10 years. All relationships between SWE (cm) and elevation (E, m) were quadratic:

Grouse Mountain

$$SWE = 1.31 - 0.0345\ E + 0.0001845\ E^2 \qquad (8.16)$$

$(r^2 = 0.99)$

Mount Seymour

$$SWE = 17.17 - 0.1511\ E + 0.0002794\ E^2 \qquad (8.17)$$

$(r^2 = 0.97)$

Hollyburn Mountain

$$SWE = 13.9 - 0.1336\ E + 0.0002752\ E^2 \qquad (8.18)$$

$(r^2 = 0.91)$

The curvilinearity was pronounced at elevations below about 500 m. Inflections generally occurred around 400 m suggesting a high frequency of rain-on-snow or other melting events at lower elevations. Above 500 m the relationship between SWE and elevation was almost linear (Eq. 8.16-8.18). Like Shaerer (1970), Claus found that in colder, drier areas, such as the Monashees, the inflection points were less pronounced and the data were more nearly linearly related.

Fitzharris (1975) analyzed snow deposition in clearings (3/4 - 3 TH in width) relative to the open (>3 TH) during 80 storms at 9 elevations on Mt. Seymour (n = 511). The regression equation representing the relationship is:

$$S(cl) = -1.3 + 1.0\ S(o) \tag{8.19}$$

$$(n = 511,\ r^2 = 0.95,\ SE = 6.9)$$

where S(cl) = snow deposition (mm) in the clearing and S(o) = snow deposition in the open (mm). For all data, the equation represents virtual identity suggesting no redistribution. Fitzharris (1975) went on to break regression equations down into the three functional elevation zones defined earlier (Ch. 8.1). The regressions were:

Snow Drift Zone

$$S(cl) = 9.4 + 0.5\ S(o) + 0.0047\ S(o)^2 \tag{8.20}$$

$$(n = 82,\ r^2 = 0.91,\ SE = 11.4)$$

Snow Zone

$$S(cl) = -2.0 + 1.0\ S(o) \tag{8.21}$$

$$(n = 186,\ r^2 = 0.95,\ SE = 6.1)$$

Wet Snow Zone

$$S(cl) = 0.0 + 0.9\ S(o) \tag{8.22}$$

$$(n = 178,\ r^2 = 0.95,\ SE = 2.9)$$

Only for the snow drift zone was there an apparent deviation from equality. This deviation was attributed to immediate redistribution of drier snow by Fitzharris (1975). We noted earlier that some of the curvilinearity was due to overload throughfall at higher elevations, possibly because of greater wind speeds (Ch. 8.1.2).

Other data relevant to the increase in snowpack with increasing elevation are summarized in Table 8.6. Clearly, there are important interactions between elevation, inter-year variability, and topographic factors. Meiman (1968) referred to early workers suggesting a decrease in snow depth at certain elevations. He stated that no reliable data support such a supposition although data exist for intermediate zones with different rates of change in snow accumulation with elevation. We first addressed the issue in Chapter 3, and rejected it as insignificant on Mt. Seymour.

It is apparent that elevation has a strong but variable influence on both snowfall in a given storm (Table 3.3) and the resulting snowpack (Table 8.6). In coastal British Columbia the within-storm relationship approximates linearity (Ch. 3.2), but the relationship between accumulated snowpack and elevation may be linear or non-linear depending on storm types for that year (Schaerer 1970, Fitzharris 1975). Our analyses of Chapter 3 suggest that the pattern will most often be curvilinear, as in Figure 8.21. Equations 8.19 - 8.22 suggest that the elevation effect dominates any potential effect of stand openings. At the elevations of ungulate winter ranges, redistribution of snow appears insignificant

Table 8.6 Effect of elevation on the snow water equivalent of the snowpack (adapted and expanded from Meiman, 1968).

Source	Location	Date	Mean increase in snow depth (inches) per 1,000' rise	Elevation range (feet)
Anderson and Pagenhart (1957)	California	April 10	20.5	6800-8600
Anderson and West (1965)	California	March 1	7-10	6000-8000
Church (1912)	California	April 27 - May 5	11.9	8200-9000
Court (1963)	California	End of March	1.8	8200-11200
Mixsell et al. (1951)	California	April 1	22	5800-7200
Curry and Mann (1965)	Alberta	?	5.4	4600-7400
Stanton (1966)	Alberta	?	15.5	5000-5500
Grant and Schlevsener (1961)	Colorado	Daily	2.7 N slope 5.7 S slope	9300-11300 10200-11300
U.S.Soil Conservation Service (1965-1967)	Colorado	Maximum pack depth	24.8[a] 15.1[b] 26.5[c] 15.3[a] 8.2[b] 12.8[c]	9600-10700 9600-10700 9600-10700 8180-9600 8180-9600 8180-9600
Gary and Coltharp (1967)	New Mexico	?	0.7 to 5.4 depending on vegetation	9900-11150
Packer (1962)	Idaho	Maximum pack depth	10.8	2700-5500
Schaerer (1970)	B.C., Seymour Mtn	?	ca.19	0-3300
	B.C., Fidelity Mtn	?	12	3200-6300
	B.C., Enderby	?	9	3900-6600
	B.C., Kimberly	?	4	3300-6200
	Alberta, Lake Louise	?	4	5200-7300
	B.C., Mt. Revelstoke	?	10	1400-6500
	B.C., Rossland	?	7	1300-5000
	B.C., Salmo	?	4	1800-4800
	B.C., Grey Creek	?	4	1800-5300

a/ = 1965, b/ = 1966, c/ = 1967

(regression coefficients near 1.0 in Eq. 8.21 and 8.22). Thus, in southern, coastal British Columbia, elevation will affect the amount of snow on the ground more than any other single factor. Predicting the extent of this effect will depend entirely on site-specific attributes which affect the vertical temperature profile, snowfall, and melt. Fitted regression equations may adequately predict snow depths at any elevation, but only for the site for which the coefficients have been derived. Furthermore, they will vary between years (Table 3.3). Generalized predictors of the elevation effect on snowpack can only be approached through a process-oriented, iterative approach such as is approximated in the model STUF (Shank and Bunnell 1982).

8.4.2 Forest Opening Effects

Forest openings tend to have a varied effect on snow accumulation largely due to the variability in the condition of deposited snow. Figure 8.22 is a simple word model depicting the major process interactions and influences that ultimately produce a given snow accumulation pattern in an opening, assuming the input of one storm event. Other assumptions of the model are discussed later.

The topography of an area, expressed as slope, aspect, and physiography, influences the wind speed and pattern of flow on a macro-scale (Ch. 4.2). The height, breadth, and penetrability of the canopy as well as the subcanopy regions

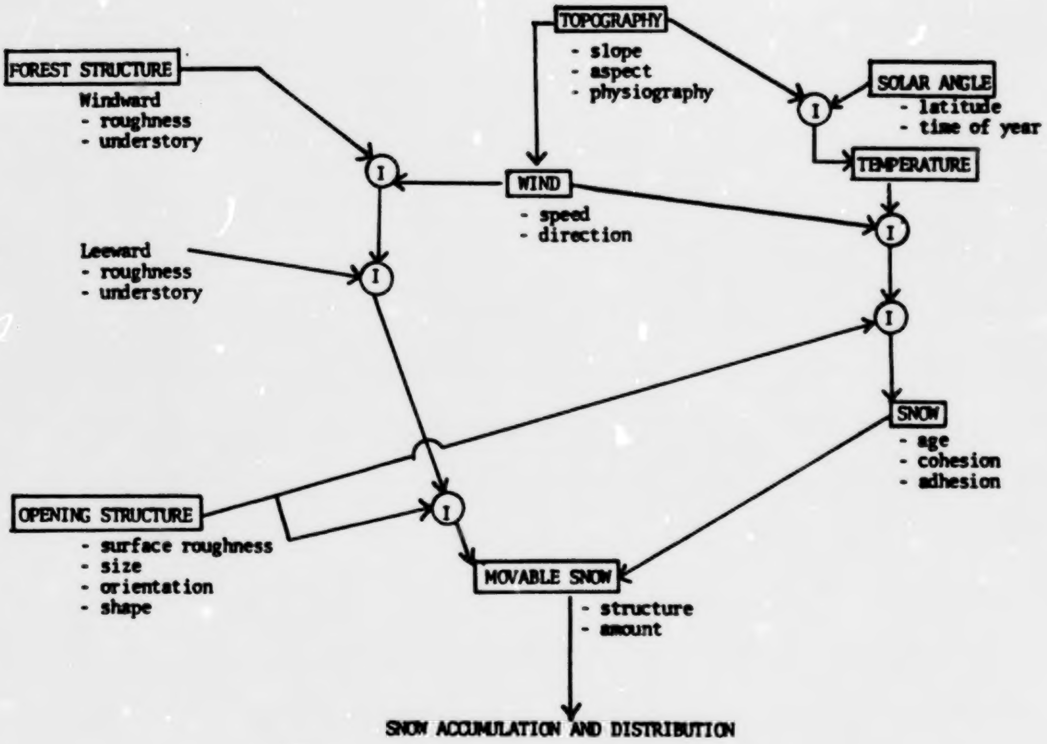
FOREST STRUCTURE
Windward
- roughness
- understory
Leeward
- roughness
- understory
TOPOGRAPHY
- slope
- aspect
- physiography
SOLAR ANGLE
- latitude
- time of year
TEMPERATURE
WIND
- speed
- direction
I
SNOW
- age
- cohesion
- adhesion
OPENING STRUCTURE
- surface roughness
- size
- orientation
- shape
MOVABLE SNOW
- structure
- amount
SNOW ACCUMULATION AND DISTRIBUTION

Figure 8.22 A word model depicting major process interactions and influences that determine snow accumulation and distribution. Interactions are depicted as $\longrightarrow$Ⓘ whereas direct influences are depicted by $\longrightarrow$.

of the windward forest determine its surface roughness. The surface roughness in turn interacts with the initial wind speed to determine the type and strength of wind flow into the leeward opening. The wind at this location then interacts with ambient temperature (which is altered by solar angle and local topography) to influence the snow characteristics both as it falls and as it sits on the leeward forest canopy. Cohesive and adhesive forces, and age of the snow determine the probability that snow will be redistributed. Depending on the penetrability of the windward forest, a recirculating eddy may form. The specific location of the eddy will be closer to the windward forest as penetrability decreases (Fig. 8.20). The leeward forest will act as a barrier of varying degrees to the recirculating flow of air. Factors determining the degree of influence of the leeward forest are its surface roughness and its proximity to the windward forest, here expressed as size of the forest opening. The surface roughness of the clearing or opening interacts with wind flow to affect its pattern of flow and velocity.

Subsequent to all the foregoing influences, the resultant windflow and velocity interact with snow once it is eroded to determine its place of deposition. Total accumulation, as influenced by redistribution, varies depending on the amount and structure of snow (Ch. 8.3.2). Accumulation cannot be expressed only in terms of redistribution and amount of free falling snow. Melt also influences the amount and pattern of accumulation in forest openings (Ch. 8.3.1). The effect of wind on snow accumulation and distribution can be viewed as a continuum. The effect is maximized in geographical locations where snow is cold and dry. The effect becomes masked as ambient temperature increases and/or as the orientation, size, and shape of the opening are such that solar angle counteracts the effect of wind through the melt process. Finally, wind's influence becomes reduced in areas where ambient temperature causes snow of a wet, heavy nature to be formed. Snow of this nature is not easily eroded and redistributed by wind, but it may be shaken off the canopy and deposited on the forest floor below the canopy periphery (Fig. 7.10).

The assumptions of the foregoing model (Fig. 8.22) are that the input at the process level is one storm event and that the geographical location (i.e., topographical variability) are the same for each storm. Also, the underlying interception surfaces cannot change in physical nature (i.e., the addition of new snow on old snow) nor can inputs such as solar angle change. These assumptions lead to constraints on the model in terms of temporal or spatial variation. Interpreting published data with respect to the model is nevertheless confusing. Studies occur across climatic gradients and published results generally reflect the ability of wind to influence the deposition or erosion of snow particles characteristic of a particular climate (Ch. 7.3). Miller (1966) pointed out that most eroded snow is deposited in openings and other areas where wind speed and flow pattern tend to favour snow deposition. Other authors consider that wind's major effect on redistribution of snow is by indirect mass transport (e.g., Church 1934, Hildebrand and Pagenhart 1954). Haupt (1979a and b) reported no apparent redistribution of snow from adjacent forests in a study in northern Idaho. Smith (1974), after watching snow placement during storms, concluded that snow is deposited evenly across openings and no differential placement was found. He attributed differential amounts of snow to differential melt at the "sunny" side of forest openings, and suggested that wind is not always the dominant process creating variability in snowpack depth. Our analyses (Ch. 7.3, 7.4, and 8.1) suggest relatively little redistribution of snow by wind on Mt. Seymour.

It is apparent that interpretation of existing literature on snow accumulation and distribution must be made cautiously because of the large temporal and spatial variation in the published data. Only general trends can be documented.

Intercepted snow in forests of coastal British Columbia usually persists in the canopy and is transported to the ground directly below the tree canopy in large clumps. Haupt (1979b) argued that the redistribution suggested by Gary (1974a) in Colorado was largely due to the climate of the Colorado Rockies. The less dense, colder snow that forms in that area would be more susceptible to wind transport and changes in wind direction such as those that occur in recirculation eddies and other turbulent air flow. Gary (1975), studied the influence wind patterns had on snow accumulation in a 1 H opening in lodgepole pine. Examination of his data indicates that maximum snow accumulation exists (as it is expected to in regions with dry, cold snow) at the zone where wind speeds and direction change abruptly due to the recirculating eddy within the clearing (Fig. 8.23). A reduction in snow water equivalent in the leeward forest suggests that snow was stolen from this area and redeposited in the opening.

Smith (1974) argued that differences in snow accumulation between forests and forest openings is not due so much to the aerodynamic effects as it is to differential melt caused by back radiation from trees. Smith recognized that wind sweeps into clearings, swirls around and leaves the opening through somewhat open stands. Whereas Gary (1975) measured differential accumulation of snow across clearings, Smith discovered no such differences during storms. Snow was deposited more evenly across the clearing because of the turbulent eddies and generally lower wind speeds. The potential for wind to blow snow underneath a canopy increases as the stand crown layer increases above the ground, and is even greater when no ground vegetation exists (Miller 1966). Gary (1975) reported that this subcanopy wind creates an even or uniform snow depth under the canopy. Fitzharris (1975) believed however, that the subcanopy wind, through eddy formation at each tree, tended to add to the formation of snow hollows around the bole of tree. Thus the subcanopy wind would contribute to an increase in variability in both depth

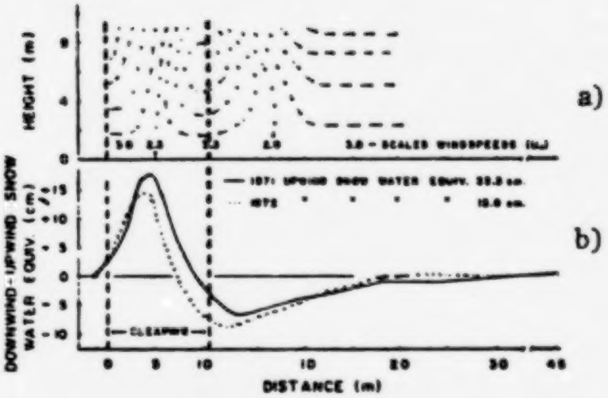

Figure 8.23 a) Measured streamlines of air flow across a clearing showing a well-developed back eddy.
b) Snow water equivalents for various distances inside and downwind of the clearing less average of upwind samples.
(from Gary 1975: 111).

and density of snow under a forest canopy. Subcanopy winds have the potential to carry snow eroded from trees, snow scoured from the ground and snow intercepted while falling, and from true throughfall (Gary 1975).

Sakharov (1949) estimated that 25 to 30 percent more snow reaches the ground under a single storied spruce stand subject to wind action than under a two-storied stand in which the understory is not shaken or exposed to wind action.

Subcanopy winds in old-growth forests at coastal British Columbia are not expected to be a significant influence on snow redistribution. Extrapolation of results from Gary (1975) would indicate that the dense undergrowth characteristic of old growth dramatically reduces subcanopy wind speed. Not only do we expect subcanopy winds to be of low velocity in old growth, but the snow deposited there will require high velocity wind before redistribution will occur because of its relatively high density (Ch. 3.3).

Table 8.7 is a collation of 15 studies involving snow accumulation and distribution in forest openings of various sizes, shapes, and orientation. The studies span a wide geographical area. Although information is sketchy it documents that the relative importance of redistribution is variable. Dietrich and Meiman (1974) were able to document the redistribution of snow in a manner relatively free from the influences of solar angle. Prevailing westerlies stole snow from the western, leeward forests and caused drifts nearer to the windward forest on northeast aspects. Melt from

Table 8.7 Reported values for the percentage increase in snow accumulation in the open over accumulation in the adjacent forest.

Source	Location	Wind direction	Plot Description					Snow water equivalent			Comment
			Size	Aspect	Orientation	Shape	Forest type	In open	In forest	%increase	
Niederhof and Dunford 1942	Colorado	-	1H[a/]	level	variable	natural	lodgepole				Maximum accumulation was in 1H open
Dietrich and Meiman 1974	Colorado	westerly	5H	S86°E		circular	lodgepole				No significant change except reduction in one plot, significant changes in distribution patterns. Wind drift and solar melt were relatively separate.
			5H	N40°E		circular	lodgepole				
			5H	N33°W		circular	lodgepole and fir				
Golding and Swanson 1978	Alberta	-	0H	level		circular	lodgepole	8.18		100	Wind redistribution and solar melt occurred simultaneously. Solar melt was believed to be relatively more important to the snow distribution pattern in the open.
			1/4H	level		circular	lodgepole	9.35		114	
			1/2H	level		circular	lodgepole	9.86		120	
			3/4H	level		circular	lodgepole	11.00		134	
			1H	level		circular	lodgepole	11.25		138	
			2H	level		circular	lodgepole	11.86		145	
			3H	level		circular	lodgepole	11.73		143	
			4H	level		circular	lodgepole	10.97		134	
			5H	level		circular	lodgepole	10.95		134	
			6H	level		circular	lodgepole	10.82		132	
Berndt 1965	Wyoming	-	8 ha[b/]	N		circular	lodgepole	18.3		122	
				S		circular	lodgepole	21.8		148	
				E		circular	lodgepole	23.4		153	
				W		circular	lodgepole	19.0		142	
			4 ha	N		circular	lodgepole	20.1		134	
				S		circular	lodgepole	20.3		138	
				E		circular	lodgepole	25.2		165	
				W		circular	lodgepole	20.8		155	
			2 ha	N		circular	lodgepole	17.8		119	Size not as critical as aspect. Wind influences not reported.
				S		circular	lodgepole	23.1		157	
				E		circular	lodgepole	23.4		153	
				W		circular	lodgepole	19.0		142	
			0 ha	N		circular	lodgepole		15.0	100	
				S		circular	lodgepole		14.7	100	
				E		circular	lodgepole		15.2	100	
				W		circular	lodgepole		13.5	100	
Rothacher 1965	Oregon	-	1.25H	S	E-W	rectangular	mountain hemlock/ true fir			135	Depths were found to be 23% greater in clearing.
Gary 1974 a	Wyoming	south-westerly	1H	S	NW-SE	rectangular	lodgepole	41.2	33.3	124	Def
Gary 1979	Wyoming	south-westerly	8-10H	N	NW-SE	rectangular	lodgepole	229	183	125)	
				N				229	175	130) 123	
				N				.91	.79	115)	
				S				218	155	141)	
				S				198	124	160) 152	
				S				102	66	155)	

Table 8.7 (continued)

Gary 1974 b	Wyoming	south-westerly	1H	S	NW-SE	rectangular	lodgepole	38.1	76.2	131	
								83.8	127.0	144	
	New Mexico	south-westerly	1H	S	E-W	rectangular	engleman	26.7	17.8	150	Shade border and sun border equal. Blowing snow counterbalanced melt Ice crust hindered snow moving.
Anderson et al. 1958	California	south-westerly	1/2H	level	-	circular	lodgepole	162.6	149.9	108	Wind redistribution patterns appear to direct accumulation patterns. Melt is less understood.
			1/2H	N				170.8	151.8	112	
			1/2H	S				141.6	131.4	108	
				E				154.3	134.0	1.15	
				W				153.7	139.1	110	
Anderson and Gleason 1959	California	-	1/2 H				true fir	120.4		137	Dense stand
Sartz and Trimble 1956	New Hampshire	-	2H				mixed hardwoods	9.8	9.1	108	
Troendle and Leaf 1980	Colorado	-	5H							133	
			6H							132	
			7H							128	
			8H							125	
			9H							122	
			10H							118	
			11H							112	
			12H							109	
			14H							100	
Golding 1982	Alberta	south-easterly	1H	E						128	Ratio is an average of 2100 openings of 3/4-1 3/4H and for 3 years of study.
Anderson 1956	California	south-westerly	1/2H	level	E-W	strip	lodgepole	50.8	40.6	125	
			1H		E-W			38.1	40.6	94	
			1H		-	square		45.7	40.6	112	
			1/2H					147.3	116.8	126	
			1H					152.5	116.8	130	
Hansen and Ffolliott 1968	Arizona	south-westerly	1 ha	N	-	block	lodgepole	36.8	25.7	143	Cut strips of a variety of widths on a southwest aspect did not produce any measurable change in snowpack characteristics.
			1/2H	S		strip		24.4	23.4	104	
			3/4H	W	E-W			26.1	15.2	172	
			1H	E	E-W			45.7	17.1	267	

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a/ Height of tree

b/ Hectare

solar radiation occurred in the northeast boundary of the opening.

Gary (1974a) could not separate the processes of wind redistribution and melt by solar radiation. In his study the processes were additive and caused a pronounced deficit of snow on the leeward portion of the opening and its adjacent forest. The leeward forest was also the area most exposed to solar radiation.

Gary (1974b), studying a similar situation in New Mexico, found the two processes to be opposing one another. The melt caused by solar radiation on the leeward border also caused a crust to be formed on the snow at that location. The crusted nature of the pack effectively reduced the ability of the recirculation eddy, caused by wind, to carry snow away from that border and into the opening. As a result the snow distribution was similar across the profile of the opening.

Golding (1982) reported that profiles of snow accumulation across circular openings of nine different sizes at James River, Alberta, were similar to those reported by Gary (1979). Despite similarities in climate and snow type, zones of excess and deficit accumulation were not as pronounced. Solar radiation apparently had the greatest influence on snow distribution patterns.

Table 8.8 shows that the ratio of snow accumulation in openings to that of the adjacent forest varies considerably from study to study. Meiman (1968) reviewed 24 studies involving differences in snow accumulating between forests as

Table 8.8 Opening sizes recommended for maximum snow accumulation.

Opening size (H)[a]	Source
2-3 H	Golding and Swanson 1978
5 H	Anderson 1970
1 H	Niederhof and Dunford 1942
2-10 H	Hoover and Shaw 1962
6-8 H	Hoover 1960
1 H	Kittredge 1953
1-2 H	Anderson et al. 1958
< 8 H	Gary 1979
5-6 H	Troendle and Leaf 1980
? openings had less snow	Dietrich and Meiman 1974
1 H	Ffolliott et al. 1965

[a]/ H = tree height of adjacent stand

compared to openings. The snow in the open was expressed as a percentage of the snow in forests and the percentage difference ranged up to 441% with most differences being less than 40%. Figure 8.24 taken from Golding (1982) depicts the relative trend that can be extracted from such data. Generally, accumulation in openings 1H and less drops off rapidly until it equals that in the adjacent forest. Rogers and Tripp (1964) found that recirculation eddies smaller than 100 m are the prevailing size in snow storms. Miller (1966) suggested that this finding reveals information about a critical dimension for openings receiving snowfall in regions where wind redistributes snow. Maximum accumulation is reached most often in openings of 2-5 H in size. However, this relationship varies considerably (Table 8.8).

Anderson (1956) provided evidence which might explain some of the variation in opening sizes which have been recommended for maximum snow accumulation. Anderson followed annual snow accumulations in three different forest conditions: i) dense lodgepole pine forest; ii) large open areas; and iii) 0.5-H strip cuts. Figure 8.25 reveals the dynamic nature of snow accumulation and retention of the two non-forested areas relative to the dense forest. Prior to April 30, both areas were similar in their ability to trap snow. Past May 20, melt became the dominant process governing snow retention and the 0.5-H opening was clearly the better opening size for retaining snow. Although Anderson's data are for late in the winter season, the same dynamic nature with respect to

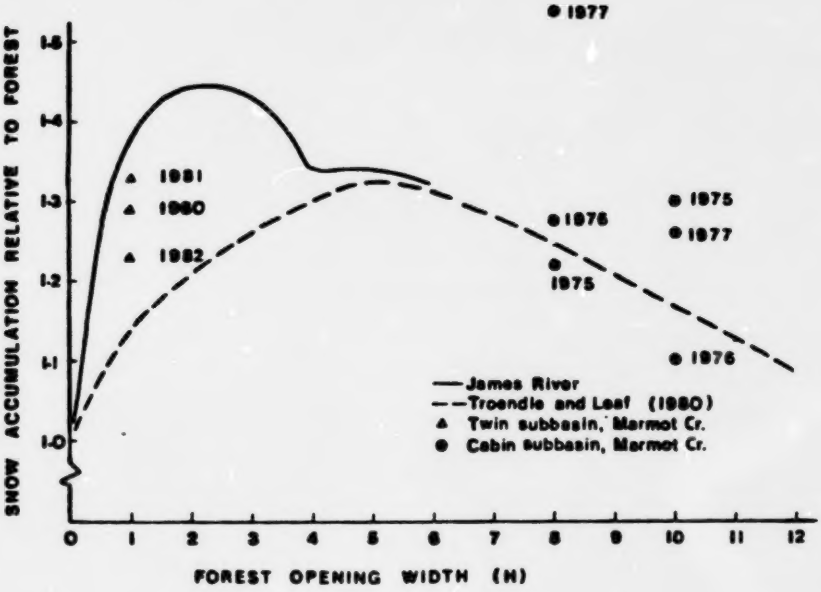

Figure 8.24 Snow accumulation in forest openings relative to accumulation in the uncut forest (from Golding 1982: 98).

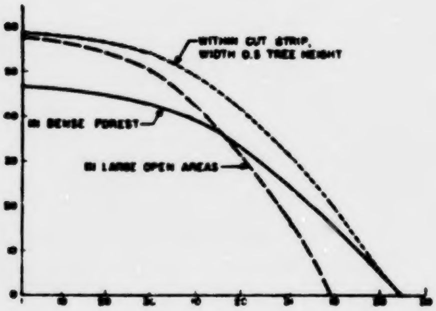

Fig. 8.25 Effect of changing climatic conditions on accumulated snow in forests, large open areas, and 1/2 H strip cuts (from Anderson 1956: 311).

ablation versus accumulation could exist any time in climates where warmer winter seasons exist (such as the low elevation areas of coastal British Columbia).

Further variation will exist depending on the slope and aspect of the prescribed opening (Ch. 4). The size of the most efficient snow trap is partially dependent on local storm conditions such as temperature and wind speed and direction during snowfall (Storr 1973). Where redistribution is held to be the primary source of increased snow accumulation, the source should be an equal-sized area of leeward forest. Troendle and Leaf (1980) recommended 5-H openings with 5-8 H leave areas in strips for maximum snow accumulation. Golding (1982) proposed that if greater than 50% of a total forested area was cut then there would be less than optimum accumulation.

To calculate the area of "leave forest", Troendle and Leaf (1980) proposed that

$$Padj = 1 + (P - 1)\ [0.5/Ao/At] \qquad (8.23)$$

where P = snow retention coefficient for open areas (taken from curve on Figure 8.24)

$Padj$ = adjusted snow retention coefficient for openings,

and, Ao = area of openings

At = total impacted area.

For southcoastal British Columbia we do not expect wind redistribution on intercepted snow to be important (Ch. 8.3.2). Opening sizes less than 1H would be unlikely to accumulate more snow than adjacent forests and are more likely to accumulate less (Tables 8.7 and 8.8, Fig. 8.24). Such small openings might experience more ablation through melt, but that is unlikely to be a large effect until late in the winter (Fig. 8.25). In short, under coastal conditions the creation of small areas to encourage forage production are unlikely to act as snow traps and may actually accumulate less snow than the forest. Under conditions where overload throughfall and mass transport are significant some small areas could accumulate more snow. The issue merits further analysis and we have suggested an approach in Appendix I.

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IV IMPLICATIONS TO MANAGEMENT AND RESEARCH

9. Towards Silvicultural Prescriptions For Winter Range

This chapter has two broad objectives:

1) to summarize major findings of this review, and

2) to summarize the most promising directions for managing stands to provide winter range.

Research directions and analyses most likely to be useful in developing stand prescriptions for winter range are presented in Appendix I.

Both objectives build on analyses of preceding chapters. Directions summarized are almost exclusively derived from material in this report. That restriction has two important implications. First, modification of forest canopies is considered only in terms of snow interception. Second, specific silvicultural regimes necessary to produce a particular canopy structure are not addressed. We treat only the 'target' structure suggested by the review.

The term 'promising directions' includes no measure of cost effectiveness. It implies only that canopies of the structure specified would prove particularly effective at intercepting snow. Many tree and stand characteristics which influence interception can be modified by silvicultural practices; there are also features over which the forester has little or no control, and which are highly unpredictable (e.g., the character of snow events). Because one objective of the review (objective 3, Ch. 1.2) was to consider how the mechanisms producing snow storms might guide geographic allocation of stand prescriptions, the prediction of snow events is also summarized. A brief summary of relevant research findings precedes comments on potential stand prescriptions.

1. In terms of the mechanisms of interception, storm types can be aggregated into two broad classes: i) those occurring in relatively calm air (wind velocities < 1 $m \cdot s^{-1}$); and ii) those in which wind velocities are greater. Storms in the latter class are more common, particularly at higher elevations where orographic effects dominate. They are less likely to produce snow at lower elevations than are the relatively calm frontal conditions sometimes associated with incursions of Arctic air (Chs. 3.1 and 3.2, Tables 3.2-3.4).

2. The first storm type occurs rarely on the coast but produces large snowfalls when it occurs. The second class occurs frequently and produces snow at variable elevations. Stand prescriptions designed to maximize interception would differ between the two broad storm

types.

3. Occurrence of the first class of storm can be treated only as probabilities, and these will be uncertain because they are products of large scale movements of air masses (Tables 3.1 and 3.2). Functional relationships with elevation can be developed for the second, more common storm type but will differ among localities (e.g., Table 3.3, Fig. 8.21).

4. The nature of the air masses involved and orographic lifting indicate that there will always be significant snowfall at higher elevations even if lower elevations receive no snow (Ch. 3). Thus, some deer will seek winter ranges at lower elevations every year.

5. An unequivocal definition of 'lower elevations' is difficult because elevation effects vary over small distances and with participating air masses. The local upper limit will vary between about 700 and 1000 m. For individual snow storms the orographic effect is locally predictable if freezing level is known (Table 3.2), but freezing levels vary with storm type (Table 3.4). In coastal mountains, snowfall generally increases monotonically with elevation (Ch. 3), but the effects of insolation and rain-on-snow events produce a curvilinear relationship between snowpack and elevation (Fig. 8.21).

Thus, conservative management of deer would suggest upper limits to winter range closer to 700 than 1000 m. Appropriate silvicultural practices may be able to raise that limit.

6. Although snowpack increases curvilinearly with elevation (Fig 8.21), the relationship appears linear at elevations where winter ranges are most useful (Eqs. 8.21 and 8.22). At those elevations, snow in small clearings is linearly related to snow in the open (Eq. 8.19). In south coastal British Columbia, elevation is the most important single factor determining snow on the ground. Aspect is secondary during the accumulation phase. Forest canopy reduces snow accumulation regardless of elevation and aspect.

7. Phenomena within individual snow storms are more predictable than the occurrence of snow storms (Chs. 3 and 4). Briefly, the most common crystal types are needles and sector plates; aggregations (snow flakes) are frequent; riming and high densities are common, producing greater mass and terminal velocities than in interior areas; both adhesion and cohesion are encouraged (Chs. 7.1 and 7.2); and wind usually accompanies the storms. Together these phenomena have several implications to future analyses and measurement techniques (App. I). The major implication to stand prescriptions is that

'redistribution' of snow is far more likely to result from overload throughfall than from stolen, eroded snow (see Ch. 8). Opening size is thus less critical to snow distribution, and the processes of interception are more important in producing snow on the ground than in drier areas. Stand openings and turbulence still will modify the pattern of snow deposition during a storm. Snow involved in overload throughfall is usually of high density and may be important in shrub burial.

8. Analyses of the review suggest the following guidelines for location of management prescriptions:

   i) Manage for crown completeness (sometimes termed crown or canopy closure, Ch. 6.1) at lower elevations (< 500-700 m); opening sizes will be important to stand prescriptions only at higher elevations where interception is less efficient (Figs. 8.1-8.6; Ch. 8.3.2). This guideline ignores forage production and the fact that small openings (< 0.5 H) could be exploited at low elevations (Fig. 8.24).

   ii) Manipulating stands to increase snow interception will be most effective in the wet zone where precipitation as rain is frequent (Fig. 8.8). Elevation of the wet zone varies within and among winters (Table 3.4) from about 200 to 600 m.

iii) Higher elevations, where redistribution and opening size are important, will be important to deer only rarely; the same is true for lower regions of the wet zone. Managing for interception on deer winter ranges should concentrate in areas that receive frequent storms of low to moderate intensity.

iv) Effectiveness of canopy completeness in different snowfall regimes can be estimated broadly. In single snow storms, efficiency of interception is about 60% when storms are ≤ 20 mm SWE (Fig. 7.14), declining to about 40% at 30 to 40 mm SWE, and continuing to decline with increasing snowfall (Figs. 7.14 and 7.15). Even with specific management to increase interception, interception efficiency is unlikely to exceed 65% at 30 to 40 mm SWE of snowfall (Table 7.2).

v) During extremely heavy snowfalls interception efficiency decreases (Fig. 7.18), and if trees have flexible, pendant branches, snow load may decrease (Fig. 7.30). The patterns of interception efficiency and accumulated snow load versus cumulative snowfall are similar (Fig. 7.12). Attempting to encourage interception by specific management of western hemlock or western red cedar in areas of high snowfall is unlikely to be rewarding.

vi) Upper asymptotes of snow load are rarely attained by stands during single snow storms (Fig. 8.7), but do influence snowpacks during a winter (Fig. 8.12). Interaction of canopy cover with maximum snowpack in the open can be used to predict snowpack in the forest (Eqs. 8.9 and 8.13, Table 9.1). Snowpack in the open is primarily a function of elevation.

vii) Available data (Figs. 8.9 and 8.12, Table 9.1) suggest that the range of effective canopy closure decreases with increasing snowfall. For example, if the target reduction in maximum snowpack was 25%, a crown closure of 20% would be sufficient if snowpack in the open was only 25 mm SWE. Crown closures of 40% would be required at 50 mm SWE and 100% at 100 mm SWE (Table 9.1). These analyses suggest the following allocation of management effort as a function of maximum snowpack (mm SWE) in the open: forage production when ≤ 30 mm; interception and forage production when between 30 and 60 mm; interception and/or forage production when between 60 and 100 mm; interception only or no management effort at all above about 100 mm.

9. The present review can offer no more specific directions regarding location of management prescriptions. We have more confidence in predicting the effects of canopy

Table 9.1 Maximum snowpack in the forest (mm SWE) as a function of canopy closure and maximum snowpack in the open (derived from Eqs. 8.9 and 8.13).

	Snowpack in open (mm SWE)				
Canopy closure (%)	25	50	75	100	125
20	19.5	43.3	68.7	95.1	122.4
40	14.0	36.5	62.3	90.2	119.7
60	8.4	29.8	56.0	85.3	117.1
80	2.9	23.0	49.6	80.4	114.4
100	0.0	16.3	43.3	75.5	111.8

closure than the effects of elevation on maximum snowpack. Successful production of appropriate crown characteristics (points 13 and 14) should raise the apparent snowpack thresholds (Table 9.1), and extend the elevation of favourable winter range.

10. At lower coastal elevations temperature conditions encourage interception, and silvicultural modifications to enhance interception should be successful. Both adhesion and cohesion are greatest at warm temperatures (Ch. 7.1), and interception increases with increasing wetness (Fig. 7.1) or increasing temperature (Fig. 7.2) in a curvilinear fashion, peaking just below freezing. Because cohesion plays a more important role on smaller surfaces (e.g., branches and needles), the effect of temperature on snow load is more strongly expressed on smaller surfaces (Fig. 7.3).

11. Warm temperatures also suggest rapid shedding of intercepted snow. Small amounts of heat can release 1 to 2 mm SWE per hour (about 20% of a moderate snow load, Figs. 7.9 and 7.10). The effect is pronounced on trees with pendant branches (Chs. 7.6.2 and 7.6.3), requiring only 1.5 langleys ($cal \cdot cm^{-2}$) to melt sufficient snow to release 1 mm SWE to mass transport. Figure 7.10 suggests a 0.9% loss of snow load for each langley of insolation even when temperatures are below 0°C. The long-term mean,

daily solar flux during winter in southcoastal British Columbia is 112 ly. The high probability of mass transport emphasizes the importance of small-scale spatial patterns in stand prescriptions. For example, openings to encourage forage production must be large enough to avoid shrub burial by mass transport.

12. Wind is common on winter ranges and has three broad effects on interception (Ch. 7.3): i) it increases adhesive and cohesive forces; ii) it removes intercepted snow by overcoming adhesion and cohesion, or by shaking; iii) it may increase the interception efficiency and snow load of complex crowns (Fig. 7.4). The relationships are complex. In trees with level whorls, snow load should increase linearly with moderate increases in wind speed (Figs. 7.3 and 7.4) and non-linearly with increasing interwhorl distance (Figs. 7.5 and 7.6). On trees with steeply sloping branches, increasing wind speed will reduce snow load and interception (Eqs. 7.7 and 7.8; Figs. 7.7, 7.10, 7.24, and 7.25). On steeply sloping crowns an increase in wind speed from 0 to 1 $m \cdot s^{-1}$ decreased snow load 25 to 30% (Fig. 7.7). At speeds above 3.5 $m \cdot s^{-1}$ there was no snow accumulation in the crown (Ch. 7.3).

13. Many attributes of individual trees influence their ability to accumulate a snow load and their interception

efficiency (% of snow falling on the horizontal area encompassed by the crown that is intercepted). The former attribute is important to the resulting snowpack, the latter is more important during single storms. Briefly -

i) Size of intercepting surface (Figs. 7.20 and 7.21): Ignoring bridging, the difference in amounts of snow intercepted by different-sized surfaces increases with increasing snowfall. At moderate snowfalls, small surfaces (e.g., needles, small branches) retain disproportionately less snow because of the time required for cohesion. Total load decreases more than interception efficiency. Effects of temperature and cohesion are more marked on narrower surfaces.

ii) Shape of intercepting surface (Figs. 7.22 and 7.23): Flat surfaces are more efficient interceptors than round surfaces, and bear larger total snow loads. The effect of shape decreases with increasing size.

iii) Slope, angle, and flexibility of branches: The amount of snow intercepted decreases with increasing slope and wind speed (Eqs. 7.15-7.21). The greater the wind speed the greater the effect of slope (Figs. 7.24 and 7.25). Total snow loads are dramatically reduced at slope angles of 60° or a height to base ratio of about 2.15 to 1 (Fig. 7.29).

Trees with flexible branches retain less snow because branch angle and height to base ratio increase with increasing snow load (Fig. 7.30). Because of wind effects (12, above), trees with horizontal whorls (initial branch angle perpendicular to the bole) catch the most snow.

iv) Interwhorl distance: The relationship is equivocal because it varies with wind speed. Under most coastal conditions, both total snow load and interception efficiency should increase curvilinearly with interwhorl distance (Fig. 7.5, Table 7.3).

14. The ideal crown for snow interception is: healthy and dense to exploit effects of surface area and roughness (Ch. 7); long, but with a height to base ratio less than 2.15:1; comprised of horizontal branch whorls that are relatively inflexible and 50 to 80 cm apart; and extends close to the ground (Figs. 7.34, 7.35, and 7.38).

15. The forester can exert control over most of the factors influencing interception by individual crowns. The most apparent controls are:

i) Species choice will influence interception. For example, Douglas-fir is 45 to 65% more efficient at interception than western white pine (Table 7.2). In coastal British Columbia, silvicultural manipulation to increase interception appears more promising for Douglas-fir than for western red cedar or western hemlock (branch flexibility appears more important than crown length when snowfall is heavy). Species choice will also constrain stocking rates and time to crown closure.

ii) Young plantations will prove inefficient interceptors primarily because of high branch flexibility and large height to base ratios. Age itself is a poor predictor of interception (Fig. 7.32). Negative attributes of age may be overcome by rapid growth rates which encourage a full crown with a favourable vertical profile of branch density (Fig. 7.34).

iii) Favourable attributes of crowns are encouraged by moderate to rapid growth; interwhorl distances greater than 80 cm may reduce interception by reducing branch density. Immediate effects of spacing or thinning will be negative (a stand attribute), but should enhance interception as crown growth responds.

iv) Pruning live branches would reduce total snow load, but increase interception efficiency (Figs. 7.38 and 7.40). The combined effect would lower total interception slightly.

16. Structure of the ideal stand is less clear because it must also accommodate forage production. Some points are evident:

    i) Immediate effects of silvicultural treatments reducing stem density will usually be negative. Thinning regimes should consider both the subsequent response of the crowns and the immediate effects of 'bridging' between trees during snow storms.

    ii) As the stand grows in height, the probability of wind redistributing snow from openings into the stand increases. Considering only interception, the ideal stand structure is multi-storied. A two-storied stand would reduce snow on the ground by 20 to 30% over a one-storied stand (Ch. 8.4.2).

    iii) A stand having few trees with deep crowns will intercept more snow than a stand having many trees with shallow crowns (Ch. 7.6.4) providing crown closure does not decline below about 40% (Table 9.1).

    iv) Prescriptions incorporating openings to provide forage should focus more on influences of radiation than on influences of wind on snow redistribution other than mass transport. Wind will be more important at higher elevations where deer are sparser; the reverse is true of radiation and heat. Snow redistribution will occur primarily during storms or rare gusting (Ch. 8.3.2). The primary interaction of wind with opening size will be through its relationships with relative humidity and melt or sublimation.

v) Openings in stands to encourage forage production should not trap snow. That restriction suggests openings of 1H or less (Table 8.5).

vi) Review of the effects of barriers suggests: The greatest effect is downwind; the higher the barrier the greater the downwind effect and the lesser the upwind effect; 40% penetrability produces the greatest downwind effect (Table 8.4). The windward forest should be managed for forage production and the leeward forest for wind recirculation (Fig. 8.20).

APPENDIX I

Promising Directions for Future Research and Analyses

This appendix utilizes findings of the review to summarize both the kinds of data and the analyses of those data that appear most promising. It thus summarizes the implications relating to measurement techniques as well as statistical analyses.

Statistical Analyses

The treatment assigns an arbitrary priority to the analyses discussed; the basis is their immediate utility, likelihood of successful completion, and time/cost involved.

1. Refinement of elevational limits to winter range: low priority. Although refinement of specific elevational limits to winter range would be useful, we rank it low in priority because of its probable poor likelihood of successful completion and the time involved to evaluate accurately the true likelihood of successful completion. Given the amount of natural variability, it is probable that individuals with local experience can do as well as all but the most detailed analyses. Furthermore, it is unlikely that sufficient data are in hand to provide more than preliminary estimates. Preliminary equations are

provided in Chapters 3 and 8; data may be available to evaluate their utility for limited areas on Vancouver Island.

2. Effects of snow density on sinking depth: very high priority. Coastal conditions (ice crystal size, shape, density, mass, and terminal velocities) produce relatively high densities of newly-fallen snow. That, coupled with frequent melting, yields snow of higher densities than is common in interior areas. These densities will either increase support of deer (lessen sinking depth and reduce energy costs of locomotion) or increase drag (increase energy costs of locomotion). If the study finds that drag is increased, spatial structure of the stand will have to consider more than interception (e.g., influences on maturity of snowpack).

3. Estimating hardness: very high priority. Coastal conditions will produce snowpacks of varying hardness; canopy structure will modify hardness. The hardness will influence energy costs of moving deer. Previous attempts to estimate hardness in a fashion meaningful to deer movement were inconclusive. A study should be designed so analyses can address the following objectives:

   i) Evaluate the precision of different methods of estimating hardness.

ii) Quantify the degree to which each method predicts sinking depth of deer ('accuracy').

iii) Document the accuracy and precision of each method under several broad canopy classes (but see also point 5 below).

iv) Provide guidelines for the appropriate methods and sample sizes to be used under different canopy types.

4. Influences of canopy cover and elevation on snowpack: High priority; analyses should be completed before stand prescriptions are developed. All data available treating the influences of canopy cover and elevation on interception during single snow storms were analysed. More data are available that treat the influences of canopy cover and elevation on snowpacks. All data should be treated before any specific stand prescription is developed for a given location. The analysis should have 3 objectives:

i) Develop predictive equations (snowpack vs elevation) using data of Table 8.6 and comparable data. Data should be stratified by 'dry' and 'wet' climates (e.g., Figs 8.8 and 8.21). The analysis differs from that of point 1 (this section), because only data for which canopy closure is known would be used.

ii) Develop predictive equations relating the influence of canopy closure to relative snow water equivalents (e.g., Eqs. 8.9 and 8.13). The analysis would document the form of the regression (e.g., Fig. 8.7), evaluate how slope changes with storm size or maximum SWE in open (e.g., Eq. 8.13), and segregate the influence of elevation by multiple regression, analysis of covariance, or simple stratification (e.g., point i above).

iii) Document the influence of measurement technique on the predictive ability of interception versus canopy closure: regressions (e.g., Table 8.3). Data of other IWIFR projects should be incorporated with data of this report. Current analyses suggest the appropriate technique will vary with canopy structure. That suggestion should be evaluated to allow efficient future monitoring.

5. Spatial heterogeneity and pattern: very high priority. Most of the analyses in this report involve single points; Chapter 8 is an exception. The lack of treatment of spatial distribution may be a major limitation to developing stand prescriptions. That is especially true if patchwise treatments are important to produce rooted forage and enhance interception concomitantly. The review asked specific questions, did not summarize coastal

samples to provide variance estimates, and cannot be used to estimate requisite sample sizes for questions not addressed. Spatial patterns will be critical to stand prescriptions; variance estimates will be necessary to develop efficient sampling regimes to monitor the effects of trial prescriptions. Both issues could be usefully addressed using data in hand. Two conditions should govern the analysis:

i) Snow variables of interest should be limited to depth, density, and hardness. Snowpacks and individual snow events should be treated separately.

ii) Initially data should be limited to those currently in hand or readily retrievable. These data include University of British Columbia data from the Research Forest, sources of Tables 8.4-8.8 and Figure 8.24, other sources collated at University of British Columbia, and data of other IWIFR projects.

To the extent that it is possible, products of the analyses should include:

i) Variance estimates for snow depth, density, and hardness segregated by broad classes of time since onset of winter, canopy cover, elevation, relative humidity, and temperature.

ii) Suggestions concerning efficient sample design including an evaluation of double sampling, cluster sampling, and transect sampling (to the extent possible from data in hand).

iii) Quantitative evaluation of the word model of Figure 8.22. Melt will be important to coastal snowpacks and was not addressed well in this review. Where possible its influence should be extracted; we expect the processes to be very complex and initial treatment could likely do no better than bound the problem. It may be sufficiently dynamic that simulation modeling is more effective than statistical analysis. Melt processes represent an important gap that cannot be treated quickly.

iv) Analysis of influences of opening sizes including the character of wind barriers (e.g., Table 8.4) and explicit recognition that openings could be constructed to trap snow or 'evade' snow. Data of Tables 8.6-8.8 and Figure 8.24 are almost all from interior conditions; judicious selection of wetter sites would more closely approximate coastal conditions.

v) A sample design for the winter of 1984-85, to address questions of spatial distribution that appear critical and poorly answered.

6. Influence of crown attributes on interception: very high priority. The conclusions drawn in the review (Ch. 7.6.4) were derived from a sparse data base. If the habitat component of IWIFR can collect estimates of interception, these conclusions should be evaluated before stand prescriptions are implemented. Evaluation would involve sampling areas from which interception estimates were obtained to document (minimally): species, broad crown form, height to live crown, crown length, and diameter of crown base. Other techniques for measuring crown closure should be incorporated into the sampling scheme (e.g., 4. iii this section). Interwhorl distance probably cannot be sampled.

Measurement Techniques

Material relevant to measurement is summarized from our review and experience; brevity does not imply complete confidence. Most of the statements are derived from review of the kinds of snow common on the coast, and the conditions under which that snow falls (Chs. 3, 4, and 5).

1. Estimates of interception derived from comparisons in the open and under the canopy will be more accurate than in drier areas (much less snow redistribution). Accuracy

will decrease with elevation (Figs. 8.1-8.6).

2. Because winds usually accompany snowstorms, there is a strong angular component to snowfall. The vertical dimension of the crown is therefore important in predicting interception (Figs. 6.2 and 6.3). Snow load is better predicted by vertically projected area ($r^2 = 0.85$) than by horizontally projected area ($r^2 = 0.65$). As expected, neither of these measures accurately predicts interception efficiency nor snow load per horizontal area of crown (Ch. 7.6.4). A simple index of crown length x crown diameter provided accurate prediction of snow load ($r^2 = 0.82$) with a low standard error (Eq. 7.36). The best predictor of snow load ($r^2 = 0.87$) and interception efficiency ($r^2 = 0.59$) was obtained from an index of crown surface area (Eq. 7.38; crown length and diameter assuming a conical shape). Age or crown length alone were not useful predictors (Ch. 7.6.4).

3. Canopy cover alone is a suboptimal measure of interception and no one technique of measuring canopy cover will be universally best (optimum represents the lowest standard error in predicting interception). The better techniques of measuring canopy cover likely will yield values of r^2 somewhat greater than 0.80 (Table 8.3). When abundant snow falls at lower elevations, conditions are likely to include calm air (Ch. 3.2) and techniques using smaller

angles (e.g., moosehorn) will predict better. Under windy conditions common at higher elevations, techniques using wider angles (e.g., convex spherical densiometers, wide angle photography) will prove better predictors (Ch. 8.2). Winds will occur in most snow storms and the latter techniques will be better techniques more frequently. This prediction should be checked (see statistical analyses, point 4. iii).

4. All studies reviewed that applied statistical analyses to differences between snow in the open and snow under the canopy made a common statistical error. Interception is calculated as a proportion of snowfall intercepted; variance of proportions must be calculated differently from variances of whole numbers. Sample sizes suggested by those studies are incorrect.

5. Given the influences of tree and stand characteristics on interception and overload throughfall the ranking of sample sizes to estimate depths will likely be old growth = 20 year old > 80 year old >> openings. Sample sizes necessary to estimate density with the same degree of accuracy will require only 20 to 30% as many samples and the ranking will differ. Both rankings will vary with elevation. Specific estimates of requisite sampling intensities are suggested under further analyses, point 5. i).